ИЗВЕСТИЯ
АКАДЕМИИ НАУК СССР
СЕРИЯ ГЕОЛОГИЧЕСКАЯ

IZVESTIYA AKAD. NAUK SSSR

SERIYA GEOLOGICHESKAYA

959 CONTENTS No. 10, October

BRIEF COMMUNICATIONS

REVIEWS AND DISCUSSIONS

TWENTY-FOURTH YEAR OF PUBLICATION

T-10006. Approved for printing September 23, 1959.
Circulation - 3800 copies. Order 3650.
Paper size 70 x 108-1/16. Paper 4-1/3. Printing sheets 11.6 + 1. Publ. sheets 13.0

Second printing office of the USSR Academy of Sciences Publishing House
Moscow, Shubinskiy per. 10.

THE DEVELOPMENT OF SCIENTIFIC RELATIONS BETWEEN GEOLOGISTS OF THE ACADEMY OF SCIENCES OF THE CHINESE PEOPLE'S REPUBLIC AND THE U.S.S.R. ACADEMY OF SCIENCES [1]

The friendly scientific relations between he geologists of two great and adjacent coun-ries, the Soviet Union and the Chinese eople's Republic, originated in pre-Revolu-onary times.

The brilliant investigations and monographs f Vladimir Afanas'yevich Obruchev, the trav-ls of N.M. Przheval'skiy, and the works of .N. Potanin contributed considerably to the ommon cause — study of the geology and eography of Central Asia.

Scientific relations grew especially after he establishment of the Chinese People's epublic. Enthusiastically and with energy, he young republic started to develop its ndustry, agriculture, and scientific institu-ons. Hardly had the state grown strong, hen the Chinese People's Republic created he Academy of Sciences; as early as 1953; t sent a large delegation to the U.S.S.R., epresenting various branches of science. Members of the geology and geography branch f the U.S.S.R. Academy of Sciences, then eaded by D.S. Belyankin, met with the cor-esponding representatives of the Chinese eople's Republic. It was then that, as scien-sts of contiguous countries constituting a geo-ogic entity, they became mutually acquanited ith both countries' geologic organizations, nd the chief problems facing them.

Since then scientific relations between both ountries have continuously expanded. Expert eologists have repeatedly come to the Soviet nion to take part in conferences covering arious branches of geology and to study geo-ogically classical areas. Huo Tih-Feng, rofessors Chjang Wen-Yu, Li P'ou and others ave been our guests.

Chinese comrades took an active part in ur wide, All-Union conferences devoted to roblems of magmatism (the first held in Moscow in 1953, the second in Tashkent in 1958) and of ore deposits (Moscow, 1959), as well as in a number of other conferences.

In connection with the compilation of the Tectonic Map of Eurasia (prepared by both Soviet and Chinese scientists) Professor Chjang-Wen-Yu, editor of the tectonic map of China, visited various structural zones of the Soviet Union, the Russian platform, the Donbas (Donets coalfield) and the Caucasus, in order to compare them with the structural zones of the Chinese People's Republic and to compile the legend of the Eurasian tectonic map. Our own experts, corresponding mem-ber of the U.S.S.R. Academy of Sciences, Yu.A. Kosygin and senior research worker M.S. Nagibin, went to China to familiarize themselves with certain controversial field data essential to the understanding of the geologic structure of the territory of the Chinese People's Republic. The first results of this common effort were discussed directly in the field.

Petrologic and metallogenic investigations of ultrabasic and basic intrusions are also an example of close contact in the field of geol-ogy. Professor Li P'on studied ultrabasic intrusions and chromite associated with them in the Urals and in other regions of the U.S.S.R. Consequently, N.V. Pavlov, of the Institute of Geology (IGEM) of the U.S.S.R. Academy of Sciences, visited the Chinese People's Republic where he, together with Chinese specialists, studied ultrabasic and basic rock localities.

These examples are sufficient testimony to the close scientific cooperation between the Chinese People's Republic and the U.S.S.R. in the solution of important geologic problems in these countries.

There are other scientifically important geologic problems which have been investigated commonly by both countries and will be, we hope, elaborated on in the future for the more fruitful development of economics and geologic knowledge.

[1] K razvitiyu nauchnykh svyazey mezhdu geologami kademii Nauk Kir i Akademii Nauk SSSR.

The laws of formation of the so-called Pacific belt structure and related magmatism and metallogeny, investigated by S.S. Smirnov, are the subject of common study by Soviet and Chinese geologists.

The construction of a chronologically absolute geologic time scale for Eurasia, based on data for igneous rocks in the U.S.S.R. and the Chinese People's Republic, also represents a problem of practical and scientific significance to be solved by geologists of both countries.

The determination of the laws of magmatic development in time and space, together with geophysical investigations, may contribute greatly to our understanding of the geologic processes which take place under plutonic conditions and, consequently, to our knowledge of the laws governing the composition and development of the earth's crust.

Such investigations require the application of comparative historical methods and the detailed and complex study of varied geologic and structural units.

There is no limit to possible joint Soviet and Chinese geologic research in this direction.

The existing collaboration of these two great nations in the maintenance of peace all over the world and the prosperity of all the working classes guarantees the success of cooperative solution of mutually important geologic problems.

METEORITES AND THE EARTH'S CRUST[1]

by

A. P. Vinogradov

I. INTRODUCTION

The data I compiled during the past two years on the distribution in igneous rocks and meteorites of the so-called rare elements and isotopes of the light elements oxygen, sulfur, and carbon, as well as the data on zonal fusion of the silicate fraction of chondrites and other silicates, could be useful in determining the origin of different types of meteorites in contrast to that of igneous rocks of the earth.

A comparison of the distribution of individual elements or their pairs (particularly of the rare elements) in the normal series of igneous rocks and in different types of meteorites which were not as thoroughly investigated, might serve as a key for understanding some of the circumstances in which meteoritic material was formed.

The isotopic composition of carbon, sulfur, oxygen (though in general they have very low dispersion) has attracted attention because of the great monotony of isotopic components of meteorites compared with the igneous rocks of the earth.

Transformations occurring in the magma or perhaps in meteoritic material have been, as a rule, investigated with a view to their products within the primary magma or some similar substance. This has so far severely restricted our concepts of the processes responsible for the distribution of matter in the earth and in meteorites. I resorted to the method of zonal fusion of chondrites and silicates. This model of fractional crystallization has led me, to a certain extent, toward a concept of "boundless" magma or parent material in the distribution process.

As a result of this research, certain ideas presented themselves concerning the mechanism of the processes of separation of meteoritic matter and the levels of cosmic gas of protoplanets, planets, or asteroids on which these processes occur.

II. SOME REMARKS ON THE COMPOSITION OF STONY METEORITES

There is a great variety of stony meteorites in terms of structure and mineralogic and chemical composition. They comprise two groups: chondrites and achondrites. The first contain chondrules, with sizes varying from one meteorite to another, and even within the same meteorite. The chondrules are commonly deformed, cemented with splinters and dust from the same chondrules. The chondrites may be divided into nonmetamorphosed, recrystallized (i.e., metamorphosed), and carbon-bearing, which contain carbon, water, and some chloritic minerals. On the average, about 12% metallic iron and about 6% troilite are contained in a chondrite mass in the form of separate grains. The chondrites consist of olivine and enstatite, bronzite, or hypersthene, with slight variations in ratio in various chondrites (Table 1). Although, in general, they resemble each other in chemical composition, the chondrites (particularly in the silicate phase) form several distinct varieties (Table 2).

The achondrites consist of two groups of meteorites which differ considerably in structure and in mineralogic and chemical composition: non-feldspathic and feldspathic. The minerals olivine and rhombic pyroxene are present, in ratios which differ in the several varieties of chondrites. It is interesting that separate chondrites are frequently found in non-feldspathic achondrites, but much less frequently in feldspathic ones. The chemical composition of the silicate phase of non-feldspathic achondrites, in comparison with that of chondrites, differs by a higher content of SiO_2 (and MgO) and a lower content of Al_2O_3, CaO, and apparently of TiO_2 and the alkalies. The ratio SiO_2:MgO in achondrites is practically the same as in chondrites:

[1] Meteority i zemnaya Kora.

Table 1

The mineralogic composition of the meteorites

Compound	Chondrites			Non-feldspathic achondrites	Feldspathic achondrites	Dunite
	According to Prior	According to Farrington	According to Wall			
Olivine	44	45,5	44.52	14.8	3,2	90,6
Pyroxene.	30	25.1	26.87	70.3	45,6	2,1
Feldspar	10	11.0	11.7	10.0	49.3	3,6
Fe + Ni	9	—	9.08	0.61	(0,5)	—
FeS	6	5,5	6,97	0.96	0,57	—

NOTE: Comma represents decimal point.

Chondrites $SiO_2:MgO$	1.60
Non-feldspathic achondrites	1.72
Carbon-bearing chondrites	1.44
Feldspathic chondrites	4.0

Feldspathic achondrites usually contain anorthite; in structure, mineralogic, and chemical composition, they correspond more than others to diabase. In chemical composition, feldspathic achondrites differ sharply from other stony meteorites in having a high content of Al_2O_3, CaO, and TiO_2.

One of the most important structural characteristics of both achondrite groups is the breccia-like texture, ranging from an admixture of splinters of mineral grains (howardites) to large splinters (chladnite, rhondite, and even some eucrites).

Of stony meteorites falling and found on earth, chondrites are by far the most numerous. Most investigators consider the genetic relations of various types of stony meteorites in the light of the different weight ratios of silicate phases — metallic iron and troilite. The relation between the nickel content in the iron phase of stony meteorites (chondrites) and the amount of iron oxide in the silicate phase is well known. As the iron oxide content in the silicate phase of chondrites increases, the quantity of metallic iron decreases, and the nickel content increases. This is the so-called "Prior rule". It is explained by the more noble nature of nickel which, in chemical equilibrium with silicate iron, chiefly enters the iron phase. The nickel content of iron sulfide (troilite) is 0.1%.

As to the processes resulting in the formation of chondrites with different iron phase content and, consequently, with a different nickel content, some investigators often fail to indicate the relationship of these processes to the general evolution of matter. Certain authors consider that these processes occur on the surface of asteroids, others find them in cosmic nebulae or inside a protoplanet. It should be mentioned, however, that meteorites, particularly stony ones, being porous bodies of low density, could have evolved only in weak gravitational fields.

III. DISTRIBUTION OF RARE ELEMENTS IN VARIOUS STONY METEORITES AND IN IGNEOUS ROCKS

The distribution of many rare elements in various phases in the process of fractional crystallization of magma can find, at present a satisfactory explanation in crystal chemistr

It is well known that certain rare element are found in definite types of igneous rocks. Thus, for instance, the content of Li, Be, Pb, Ca, Ga, Sr, Y, Tr, Nb, U, Th, and other elements tends to increase in acid rocks, whereas the greatest content of Cr, Ni, Pt, Co, V, etc., is usually found in basic and ultrabasic rocks. This example shows that the distribution of rare elements may be correlated with the presence and amount of SiO_2, etc., in igneous rocks.

Thus, the distribution and association of a number of chemical elements in igneous rocks points somehow to their origin and to the conditions of concentration in the given rock. The main difficulty here lies in the paucity of precise data concerning rare elements in meteorites. I shall deal therefore only with the most reliable information. We should note the regularity of rare element content within the several types and groups of stony meteorites. However, the distribution of alkali elements is more peculiar. According to reliable data, the potassium

Table 2

Average chemical composition of the silicate phase of stony meteorites

Type of stony meteorites	Number of cases	SiO_2	MgO	FeO	Al_2O_3	CaO	TiO_2	MnO	K_2O	Na_2O	Cr_2O_3	P_2O_5	Fe	Ni	Co	FeS
Non-feldspathic achondrites	15	52,56	30,47	11,45	1,09	1,20	0,12	0,39	0,11	0,36	0,83	0,10	2,68	0,17	0,0	0,96
Feldspathic achondrites	25	48,65	9,87	16,31	11,71	10,39	0,50	0,47	0,27	0,83	0,40	0,10	1	0,1	0,0	0,57
Chondrites	94	47,04	29,48	15,40	3,09	2,41	0,14	0,31	0,21	1,21	0,45	0,26	11,76	1,34	0,08	5,73
Carbon-bearing chondrites	11	37,53	26,03	25,75	3,05	2,66	0,13	0,34	0,17	1,22	0,54	0,38	0,0	0,0	0,00	10,61

NOTE: Comma represents decimal point.

content in chondrites reaches an average of $8.5 \times 10^{-2}\%$. Feldspathic achondrites contain half this amount of potassium. Non-feldspathic achondrites have the most variable and lowest content of potassium of all stony meteorites. It is true that among iron stony meteorites, the potassium content in olivine is as low as $1 \times 10^{-3}\%$. In igneous rocks of the earth, such a low potassium content is encountered in ultrabasic rocks, particularly in dunites. I stress it because dunites play a certain role in a concept which I shall describe later. However, one cannot accept the data of V. Kholik and L.Kh. Arens on the potassium content of dunites being less than $1 \times 10^{-3}\%$. Our determinations show that the potassium content in dunites is, on the average, less than $2.8 \times 10^{-2}\%$; yet, as noted in Table 3, it is still considerably lower than in chondrites.

The determinations of rubidium content given below are quite reliable; therefore, in our case, the ratio K:Rb in dunites will be an inexplicable exception to the general behavior of this pair of chemical elements. The content of all other rare alkalies (Li, Rb, Cs, and Na) in stony meteorites is parallel to that of potassium.

The content of all alkalies decreases, starting from chondrites and down to non-feldspathic achondrites with highest content of SiO_2; this being contrary to the rule according to which alkali increases with the increase of SiO_2 in the normal series of igneous rocks in the earth's crust. Since I am not yet dealing with the genetic relations between various types of stony meteorites, one may doubt at this stage the validity of placing the various stony meteorites within the range of variation of their SiO_2 content.

Owing to the low strontium content, the ratio Ca:Sr in all stony meteorites and dunites fluctuates from 1,000 to 2,000, whereas the ratio in igneous rocks of the earth's crust fluctuates from 50 to 200.

The content of barium is also very low. Data concerning barium are not very reliable due to insufficient sensitivity of methods of determination (flame photometry and spectroscopy). Its smallest quantity is found also in non-feldspathic achondrites and in dunites. The Ca:Sr:Ba ratios in stony meteorites conform with the peculiar process of differentiation in stony meteorite materials, and with its very shallow depth compared with igneous rocks. This can be seen in Table 4 and in the Figures 1-4, where variation curves are plotted on the same scale for a series of elements in stony meteorites and in igneous rocks.

The Al:Ga ratio may serve as an even

Table 3

Potassium content in dunites

Number of sample	Region	Massif	K in %
1	Chinese People's Republic	Yuyshigoy Institute of Geologic Sciences of the Chinese Academy of Sciences	5.2×10^{-2}
2	" " "	Nan'Shan' Institute of Geologic Sciences of the Chinese Academy of Sciences	1.2×10^{-2}
3	" " "	Same	6.3×10^{-2}
1002	Polar Ural (U.S.S.R.)	Syum-Ken	3.6×10^{-2}
1004	" " "	Ray-Iz	0.35×10^{-2}
1005	" " "	Syum-Ken	1.1×10^{-2}
1007	" " "	Ray-Iz	2.1×10^{-2}
1009	Northern Siberia (USSR)	Sabyda, Gulin Intrusive	3.7×10^{-2}
1023	Polar Ural (U.S.S.R.)	Sobsko-Voykarsky massif	1.8×10^{-2}
1025	" " "	" " "	1.7×10^{-2}
1044	Central Urals (U.S.S.R.)	Tagil massif	4.0×10^{-2}
		Average	2.8×10^{-2}

more striking indication. The gallium content is undoubtedly at a minimum in non-feldspathic achondrites and dunites. On account of the high aluminum content of feldspathic achondrites, one would have expected a considerable rise in gallium content also, in comparison with chondrites. But, on account of the high gallium content in the iron phase of chondrites, it is very difficult to determine the gallium content to the order of $10^{-4}\%$ in its silicate phase. Yet, inasmuch as the gallium content is higher in feldspathic than in non-feldspathic achondrites, one must assume that the rule which regulates the constancy of the Al:Ga ratio in stony meteorites is applicable here also. This is a very important question, the answer to which will point to the way in which the aluminum content increases in feldspathic achondrites.

The U:Th ratio in stony meteorites, as well as in igneous rocks, fluctuates within a range of about 5. The absolute uranium and thorium content of stony meteorites is still causing disagreement among various authors. I will not dwell on this question, though a great amount of literature has been devoted to it and though it has an important bearing on the estimate of the thermal balance of the earth, and on other points. I will only remark that, inasmuch as the radioactivity method based on the barium and xenon yields gives similar results, the uranium content in chondrites must be greater than 1 to $2 \times 10^{-6}\%$. The quantity of uranium and thorium does not increase in stony meteorites with the increase of SiO_2 (e.g., in non-feldspathic achondrites) but rather decreases, as (according to uranium determination) in Johnston and other achondrites. Undoubtedly, the uranium content in dunites is lower than in chondrites.

The order of magnitude of halogen content in stony meteorites is the same as in igenous rocks. The lowest quantity of halogens is contained in dunites.

Moreover, if we also take into account some data on distribution of yttrium, zinc, lead, and molybdenum in meteorites, as well as the qualitative data for many other rare elements, as compared with igneous rocks, then the following conclusion may be drawn. In general, stony meteorites contain much smaller amounts of terrestrial rare elements than do igneous rocks. In stony meteorites with higher SiO_2 content, the concentration of alkalies decreases; this is the contrary of what happens in igneous rocks. Such a loss of alkalies may be due to the high temperatures of the processes by which the meteoritic material was differentiated into the various types of stony meteorites. But this process did not reach great depth, that is, the degree of separation of individual components was insignificant. Therefore the process of separation into various types of stony meteorites was not similar to, or rather, not identical with the processes taking place in the earth's crust, such as liquefaction, fractional crystallization, hydrothermal action, and so on.

Table 4

The alkali content, alkaline elements and gallium in stony meteorites and dunites (in %)

Name	K	Rb	Cs	Li	Sr	Ba	Ga*	Author
				Chondrites				
Average of 21 determinations	$9\cdot10^{-2}$	$9\cdot10^{-4}$	—	$2,7\cdot10^{-4}$	$1\cdot10^{-3}$	$8\cdot10^{-4}$	—	L. H. Ahrens, W.H. Pinson, 1952 to 1953
" " " "	—	—	—	$3,5\cdot10^{-4}$	—	—	—	L.W. Strock, 1936
" " " "	$8,5\cdot10^{-2}$	$5,2\cdot10^{-4}$	—	—	—	—	—	G. Edwards, H.C. Urey, 1955
Saratov	$8,3\cdot10^{-2}$	—	—	—	—	—	$5\cdot10^{-4}$	Our determinations
Holdbrook	$8\cdot10^{-2}$	$8\cdot10^{-4}$	$4,9\cdot10^{-5}$	—	—	—	—	B.M. Gordon, L. Friedman, 1957
Beardsley	—	—	$9\cdot10^{-5}$	—	—	—	—	" " " " " "
Average of 3 determinations	$8,1\cdot10^{-2}$	$2,6\cdot10^{-4}$	$9\cdot10^{-6}$	—	—	—	$3\cdot10^{-4}$	H. Onishi, E.B. Sandell, 1957
				Feldspathic achondrites				
Stannern	$5,5\cdot10^{-2}$	$2,5\cdot10^{-4}$	—	$3\cdot10^{-4}$	$6\cdot10^{-3}$	—	—	Our determinations
Yurtuk	$5\cdot10^{-2}$	$6\cdot10^{-4}$	—	$2,2\cdot10^{-4}$	$4\cdot10^{-3}$	—	$3\cdot10^{-4}$	" "
Pazamont	—	—	$2,8\cdot10^{-5}$	—	—	—	—	B.M. Gordon, L. Friedman, 1957
Average of 10 determinations	$4,7\cdot10^{-2}$	—	—	—	—	—	—	G. Edwards, H.C. Urey, 1955
				Non-feldspathic achondrites				
Norton County	$3\cdot10^{-2}$	$4\cdot10^{-4}$	—	$9\cdot10^{-5}$	$7.5\cdot10^{-4}$	—	$1\cdot10^{-4}$	Our determinations
Staroye Pes'yanoye	$7\cdot10^{-2}$	$7\cdot10^{-4}$	—	$1,5\cdot10^{-4}$	$7\cdot10^{-4}$	—	$1\cdot10^{-4}$	" "
Johnstown	$1\cdot10^{-3}$	—	—	—	$3\cdot10^{-4}$	$5\cdot10^{-4}$	—	L.H. Ahrens, W.H. Pinson, 1952 to 1953
Olivine from Pallasite	$1\cdot10^{-3}$	—	—	—	$1\cdot10^{-4}$	$7\cdot10^{-4}$	—	L.H. Ahrens, W.H. Pinson, 1952 to 1953
				Dunites				
Polar Ural, Syum Ken (1005)	$1\cdot10^{-2}$	$1,3\cdot10^{-4}$	—	$1\cdot10^{-4}$	—	—	$3\cdot10^{-4}$	Our determinations
Northern Siberia, Sabyda-Gulin (1009)	$3,7\cdot10^{-2}$	$2,4\cdot10^{-4}$	—	$n\cdot10^{-4}$	$3.5\cdot10^{-3}$	—	$3\cdot10^{-4}$	" "
Northern Ural, Tagil (1044)	$4\cdot10^{-2}$	$2\cdot10^{-4}$	—	$6\cdot10^{-5}$	$2\cdot10^{-4}$	—	$3\cdot10^{-4}$	" "
Average of 4 determinations	$1\cdot10^{-3}$	$2\cdot10^{-4}$	—	$3\cdot10^{-5}$	$1\cdot10^{-4}$	$6\cdot10^{-4}$	—	L.H. Ahrens, W.H. Pinson, 1953
" " 5 "	—	—	—	—	—	—	$1\cdot10^{-4}$	A.A. Borisenok (not publ.)
" " 11 "	$2,8\cdot10^{-2}$	$1,7\cdot10^{-4}$	—	$7\cdot10^{-5}$	$4,5\cdot10^{-4}$	—	—	Our determinations
" " 13 "	$1\cdot10^{-3}$	—	—	—	—	—	—	W. Holyk, L.H. Ahrens, 1953

* In the silicate phase.

NOTE: Comma represents decimal point.

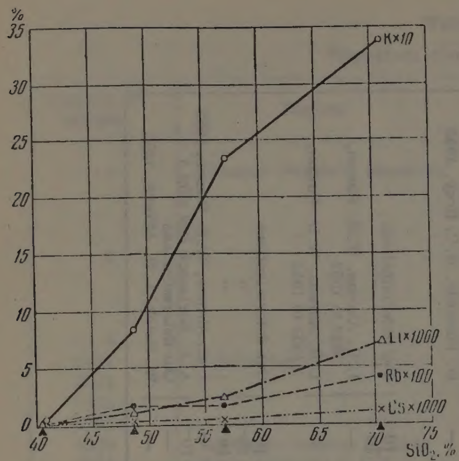

FIGURE 1. Variation diagram of alkalis in a series of igneous rocks, as a function of SiO_2 content.

[8], the geochemical importance of these radioactive isotopes of the earth's crust. Here I deal only with the stable isotopes, and mainly with the light nuclides.

The isotopic ratio in a number of chemical elements found in rocks, ores, and minerals points to the conditions of their origin and,

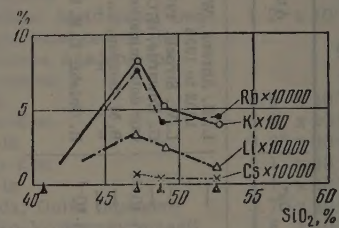

FIGURE 2. Variation diagram of alkalis in stony meteorites, as a function of SiO_2 content.

In particular, the fact that dunites, in comparison with achondrites, contain most of the rare elements by at least 1/2 to 1 order lower, permits an assumption that dunites are residual rocks from the fusing-out of the light fraction (of the earth's crust under terrestrial conditions) from the mantle material, corresponding in composition to chondrites.

4. ISOTOPIC COMPOSITION OF IGNEOUS ROCKS AND METEORITES

To date, nearly fifty natural radioactive isotopes have been found on earth apart from stable isotopes (not counting the decay series of uranium and thorium). I have already discussed to some extent, in my 1957 lecture first of all, to the temperature of their formation and to the regenerative or oxidizing character of the environment. From thermodynamic considerations one would expect the greatest effect of isotopic separation, in natural processes, among the light atoms. Disturbances of isotopic ratios occur, in natural processes, mainly by isotopic exchange which takes place in gases, liquids, and fused matter. A long-lasting state of chemical equilibrium under geologic conditions ensures the completion of full isotopic exchange. The equilibrium distribution of isotopes, in general, brings about their unequal distribution in aggregates, i.e., phases, minerals. However, the equilibrium constant in such exchange approaches 1 with increasing temperature, that is, the separation of isotopes becomes,

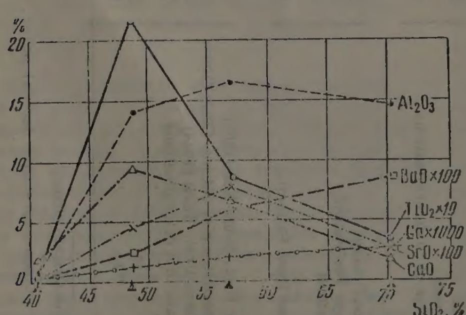

FIGURE 3. Variation diagram of Al_2O_3, Ca, BaO, SrO, TiO_2, CaO in a series of igneous rocks, as a function of SiO_2 content.

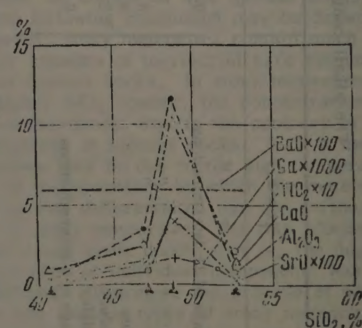

FIGURE 4. Variation diagram of Al_2O_3, Ca, BaO, SrO, TiO_2, CaO in stony meteorites.

as a rule, improbable. It is known, in part, from the literature on the subject, that dispersion of the isotopic content of carbon, oxygen, sulfur, hydrogen, and several other light elements occurs, in a way which depends on the source material. However, the number of observations were insufficient for drawing any conclusion. Moreover, the authors used different standards, which did not make things easier. Therefore, we carried out systematic determinations of isotopic ratios of the above-mentioned elements in the principal igneous rocks and in meteorites of various types. The margin of error in our mass spectrometric determinations of isotopic ratios, notably of sulfur, oxygen, and carbon, was ± 0.01%. We described our methods of investigation elsewhere [10].

I can dwell here only on those results which bear directly upon the problem. Thus, we found that sulfur in iron, iron-stony, and stony meteorites is quite the same, as to isotopic content $S^{32}:S^{34}$ = 22.20, which indicates the regenerative character of the formative environment of meteorites (Tables 5 and 6).

Sulfur of volcanic origin shows very little range in isotopic ratio, and approximates in this respect the isotopic content of meteoritic sulfur (Table 7). A more pronounced dispersed distribution of isotopes of sulfur occurs in igneous rocks, especially in magmatic sulfides. The isotopic sulfur content in basic rocks, and in particular granites, differs from that of meteoritic sulfur (Table 8).

Thus, the isotopic sulfur content in all meteorites — iron, iron-stony, and stony — as well as sulfur from volcanos and sulfur of dunites is very similar or even identical. But the basic and acid igneous rocks contain an isotopic sulfur content which differs noticeably from that of meteoritic sulfur. However, as a rule, a decrease of the $S^{32}:S^{34}$ ratio occurs in rocks ejected on the surface; in other words, they become relatively enriched in S^{34}. A still greater range in $S^{32}:S^{34}$ is observed in the biosphere under highly oxidized conditions (Table 9).

Quite the same tendency — increasing dispersion of isotopic oxygen content — is also found in igneous rocks from deep-lying rocks (dunites) to more superficial ones (granites). This is especially striking with the isotopic oxygen content in dunites as compared with that in granites (Tables 10, 11, and 12).

If granites alone deviated somewhat from the norm in isotopic oxygen content, this could be explained by the high molecular quartz SiO_2 content in acid rocks which, like the molecule $CO_2{}^{18}$, is considerably enriched in O^{18}. However, as shown in Table 11, a certain dispersion of the isotopic content of O_2 occurs alos in basic rocks. In ultrabasic rocks, particularly dunites, the isotopic content of $O^{16}:O^{18}$ is close to the same ratio in stony meteorites (Table 13).

The determination of isotopic oxygen content in chondrites, in various achondrites, and in carbon-bearing meteorites has shown that this ratio in chondrites and achondrites (nonfeldspathic and feldspathic) is the same and cannot be distinguished. Carbon-bearing

Table 5

The isotopic composition of sulfur of iron meteorites

Meteorites	S^{32}/S^{34}	Author, Year
Iron:		
Avgustinovka (U.S.S.R.)	22.10*	A.V. Trofimov, 1949
Sikhote-Alin "	22.20*	" " " "
" " "	22.20	A.P. Vinogradov et al., 1956.
Canyon Diablo (U.S.A.)	22.20	D. Maknamara and his co-workers, 1956
Dashen (U.S.A.)	22.23	" " " " " "
Toluca (Mexico)	22.24	" " " " " "
El-Trebol (Argentina)	22.22	" " " " " "
Gocharus (Southwestern Africa)	22.22	" " " " " "
Iron-stone:		
Brenham (U.S.A.)	22.23	" " " " " "

* Correction within experimental error.

Table 6

The isotopic composition of sulfur of chondrites and achondrites

Meteorites	S^{32}/S^{34}	Author, Year
Chondrites:		
Saratov (U.S.S.R.)	22.25*	A.V. Trofimov, 1949
Grosslibental	22.10*	" " " "
Vengerovo	22.20	A.P. Vinogradov et al., 1957
Yelenovka	22.20	" " " " " "
Zhovtnevyy farmstead	22.20	" " " " " "
Kunashak (gray variety) (U.S.S.R.)	22.20	" " " " " "
Kunashak (black variety)	22.20	" " " " " "
Nikol'skoye " "	22.20	" " " " " "
Orlovka " "	22.20	" " " " " "
Okhansk " "	22.20	" " " " " "
Pervomayskiy farm	22.20	" " " " " "
Union County (U.S.A.)	22.20	J. Maknamara and his co-workers, 1952
Achondrite-Chladnites:		
Staroye-Pes'yanoye (USSR)	22.20	A.P. Vinogradov et al., 1957

* Correction within experimental error.

Table 7

Isotopic composition of volcanic sulfur

Location	S^{32}/S^{34}	Author, Year
Grand-Soufrière Crater, Guadeloupe, West Indies	22.16	J. Noetclin, 1952
Same	22.18	" " "
Same	22.10	" " "
Naguen Valley, Vanna Lava, New Hebrides Isl.	22.30	" " "
Porte de l Enfer, Pozzuoli, (Italy)	22.30	" " "
Klyuchevskaya mound, Kamchatka (U.S.S.R.)	22.10	A.V. Trofimov, 1949
Kunashir Island, Kuril' Islands, Mendeleev volcano	22.20*	A.P. Vinogradov and co-workers, 1957
Same	22.20**	" " " " "
Same	22.20***	" " " " "
Stromboli (Italy)	22.200	" " " " "
" "	22.250	" " " " "
Naples "	22.20	" " " " "
Isan volcano (Japan)	22.184	" " " " "
Krasuvik (Iceland)	22.24	" " " " "

* Exhalation sulfur.

** Vein sulfur.

*** Melanterite, fusion zone.

Table 8

Isotopic composition of sulfur of igneous rocks

Rocks, location of sample	S^{32}/S^{34}	Author, Year
Granite; Caucasus, Tyrny-Auz	22.009	A. P. Vinogradov, 1957
Same; Susamyrsk	21.556	" " " "
Basalt; Kamchatka	22.164	" " " "
Same; Carpathians	22.192	" " " "
Pyroxenite; Aldan	22.20	" " " "
Dunite; Ural, Ray-Iz	22.20	" " " "
Same	22.20	" " " "
Dunite; Ural Syum-Keu	22.20	" " " "
Same; Ray-Iz	22.20	" " " "

Table 9

Average data for the isotopic abundance ratios of sulfur in igneous rocks, ores, and meteorites

Name of material	Number of occurrences	S^{32}/S^{34}
Stony meteorites	13	22,20
Iron meteorites	10	22,20
Volcanic sulfur	15	22,21
Ultrabasic rocks (Dunites)	4	22,20
Basalts	2	22,178
Granites	2	21,780
Magmatic sulfides	7	22,138
Hydrothermal sulfides	32	22,148
High-temperature sulfides		21,885
Sea sulfides	13	21,75

Table 10

Isotopic composition of oxygen in acid rocks

Object of investigation	Place of collection	Deviation from standard in %	Absolute value of O^{16}/O^{18}	Mean absolute value of O^{16}/O^{18}
Granite	Ukraine, quarry Podstepnoye	—0,36	—	488,8
"	Ukraine, Krovorogzh'ye	—0,36	—	488,8
"	Ukraine, Shpola Prudyanskiy quarry	—0,28	—	488,4
"	Ukraine, Kirovgrad, Sugakleya River	—0,03 —0,02	487,1 487,1	487,1
Trachytic granite	Ukraine, Novoukrainka	—0,04 —0,05	487,2 487,2	487,2
	Average	—0,16	—	488,0
Standard quartz	Polar Ural, Neroyka deposit	—	—	487±1

NOTE: Comma represents decimal point.

Table 11

Isotopic composition of oxygen in basic mountain rocks

Object of investigation	Place of collection	Deviation from standard in %	Absolute value of O^{16}/O^{18}	Mean Absolute value of O^{16}/O^{18}
Basalt	Kamchatka, Klyuchevskaya volcano, Kirgurich	-0.48 -0.45	489.3 489.2	489.25
"	Same, Bilyukay crater, 1938 eruption	-0.38 -0.37	488.8 488.8	488.8
"	Kamchatka, Tolbachik	-0.70 -0.70	490.4 490.4	490.4
Gabbroic diabase	Eastern Siberia, Mogdy River, tributary of the Vilyuy (sample no. 296/55)	-0.41 -0.41	489.0 489.0	489.0
Taxitic diabase	Noril'sk (sample no. 7/2458	-0.41 -0.42	489.0 489.0	489.0
Trachytic diabase	Eastern Siberia, Vilyyuy River (upper part), (sample no. 255/55)	-0.42 -0.42	489.0 489.0	489.0
	Average	-0.46	—	489.2

Table 12

Isotopic composition of oxygen in ultrabasic rocks

Object of investigation	Place of collection	Deviation from standard in %	Absolute value of O^{16}/O^{18}	Mean Absolute value of O^{16}/O^{18}
Dunite	Polar Ural, Syum-Key massif	-0.78 -0.76 -0.76	490.8 490.7 490.7	490.7
"	North Ural, Kytlym, Konzhakovskiy stone	-0.65 -0.72	490.2 490.5	490.35
"	Northern Siberian platform, Sabyda, Gulin intrusion	-0.60 -0.54	489.9 489.6	489.75
"	Polar Ural, Syum-Key massif	-0.57 -0.56	489.8 489.7	489.75
"	Polar Ural, Ray-Iz massif	-0.46 -0.46	489.2 489.2	489.2
	Average	-0.62	—	490.0

Table 13

Isotopic composition of oxygen in iron-stone meteorites (pallasites) *

Name of meteorites	Place of collection	Deviation from standard in %	Absolute value of O^{16}/O^{18}	Mean absolute value of O^{16}/O^{18}
Pallasite iron	RSFSR, Medvedeva village, Krasnoyarsk kray; finding of 1749	—0,74 —0,74	490,6 490,6	490,6
Bragin	BSSR, Polesskaya oblast', Bragin region; finding of 1937	—0,69 —0,70	490,4 490,4	490,4
Marjalhafi	Finland, fall of June First, 1902	—0,68 —0,66	490,3 490,2	490,25
	Average	—0,70	—	490,4

* Silicate phase

Table 14

Isotopic composition of oxygen in chondrites

Name of meteorites	Place of collection	Deviation from standard in %	Absolute value of O^{16}/O^{18}	Mean absolute value of O^{16}/O^{18}	Remarks
Saratov	RSFSR, fall of 6 October 1918	—0,75 —0,72	490,6 490,5	490,55	Non-magnetic fraction
Zhovtnevyy farm	USSR, Stalinskaya Oblast', fall of 9 October 1938	—0,67 —0,67	490,3 490,3	490,3	Olivine fraction
Yelenovka	USSR, Stalingrad Oblast', St. Yelenovka; fall of Oct. 17, 1951	—0,65 —0,64	490,2 490,1	490,15	Non-magnetic fraction
Nikol'skoye	RSFSR, Moscow Oblast', Solnechnogorsk; fall of 6 March 1954	—0,67	490,3	490,3	Non-magnetic fraction
	Average	—0,68	—	490,3	

NOTE: Comma represents decimal point.

chondrites or achondrites are an exception. Their isotopic oxygen conten deviates considerably, tending to increase in O^{18} (Tables 14-16).

One should note that carbon-bearing meteorites are also exceptional in regard to the isotopic content of other light elements such as carbon and hydrogen. The reason for this is not yet quite clear. It appears from Table 17 that a tendency toward the same isotopic ratios exists for O_2 as for sulfur in igneous rocks, namely, an increase in O^{18}, especially marked in acid rocks and in rocks of the biosphere.

The isotopic ratios $C^{12}:C^{13}$ in igneous rocks and in meteorites are very similar.

Table 15

Isotopic composition of oxygen in achondrites*

Name of meteorites	Place of collection	Deviation from standard in %	Absolute value of O^{16}/O^{18}	Mean absolute value of O^{16}/O^{18}
Stannern, feldspathic eucrite	Czechoslovakia; fall of 22 May 1808	—0,78 —0,80	490,8 490,9	490,85
Chervonyy Kut, feldspathic eucrite	USSR, Talalayevka region, Sumsk Oblast', fall of 23 June 1939	—0,56 —0,58	489,7 489,8	489,75
Vavilovka, feldspathic howardite	USSR, Maksimovka village, Skadoysk region, Kherson Oblast'; fall of 19 June 1876	—0,67 —0,66	490,3 490,2	490,25
Staroye Pes'yanoye non-feldspathic chladnite	RSFSR, Vargashin region, Kurgan Oblast'; fall of 2 October 1933	—0,68 —0,68	490,3 490,3	490,3
Norton County, non-feldspathic Bustide	Nebraska; fall of 18 February 1948	—0,50 —0,50	489,4 489,4	489,4
	Average	—0,64	—	490,1

* Non-magnetic fraction

Table 16

Isotopic composition of oxygen in carbon-bearing meteorites*

Name of meteorites	Place of collection	Deviation from standard in %	Absolute value of O^{16}/O^{18}	Mean absolute value of O^{16}/O^{18}
Novyy Urey, uralite	RSFSR, Karamzinka village, Ardatov region, Gor'kovakaya Oblast'; fall of 4 September 1886	—0,40 —0,38	488,9 488,8	488,85
Migei, carbon-bearing chondrite	USSR, Pervomaysk region, Odessa Oblast'; fall of 21 June 1889	—0,41	—	489,0
Groznaya, carbon-bearing chondrite	North Osetiya, Mikentskaya stanitsa; fall of 28 June 1861	—0,39	—	488,9
Staroye Boriskino, carbon-bearing chondrite	RSFSR, Chkalov Oblast'; fall of 20 April 1930	—0,40	—	488,9
	Average	—0,40	—	488,9

* Non-magnetic fraction

Table 17

Isotopic composition of oxygen in meteorites and igneous rocks

Object of investigation	Number of determinations	Deviation from standard in %	Absolute ratio of O^{16}/O^{18}
Meteorites			
Silicate phase of pallasites	6	—0,70	490,4
Chondrites	7	—0,68	490,3
Feldspathic achondrites	8	—0,66	490,3
Non-feldspathic achondrites	4	—0,59	489,9
Carbon-bearing chondrites	5	—0,40	488,9
Igneous rocks			
Dunites	10	—0,62	490,0
Basic rocks (basalts)	6	—0,42	489,2
Acid rocks (granite)	7	—0,16	488,0
Standard (quartz)	—	—	487,1

espite much information on the $C^{12}:C^{13}$ tios for igneous rocks and meteorites, ese data are hardly comparable. But cording to certain authors, there is a rceptible tendency toward increased C^{13} ntent in the more acid rocks of the earth's rface. This can be clearly seen in our ` ries of comparative determinations in igne-s rocks and meteorites (Tables 18-19).

In the biosphere, a considerable separation carbon isotopes occurs, due to disposition of eat masses of $CaCO_3$, the molecule of which s a maximum equilibrium content of C^{13}.

The $C^{12}:C^{13}$ ratio in iron meteorites often gives a greater dispersion of $C^{12}:C^{13}$ than does the isotopic content of stony meteorites.

The dispersion of the isotopic carbon content of iron meteorites is due to the formation, at high temperatures, of the carbide Fe_3C (so-called cohenite). The isotopic ratio $C^{12}:C^{13}$ of this Fe_3C phase, as shown by H. Crag, is 90.5:1, whereas for the diffused carbon in the Canyon Diabolo iron meteorite this ratio is equal to 89.45:1. In isotopic ratios, carbon in dunites is even closer to meteoritic carbon.

Table 18

Average data on C^{12}/C^{13} for rocks and meteorites according to R. Craig, F. Birch, A.V. Trofimov

Name of material	Number of determinations	Dispersion C^{12}/C^{13}	Mean C^{12}/C^{13}
Aragonite	92	88,1—89,5	89,0
Iron meteorites	13	89,6—91,8	90,6
Stony meteorites	26	89,6—91,2	90,6
Igneous rocks	13	90,3—91,2	90,8

NOTE: Comma represents decimal point.

Table 19

Isotopic composition of carbon, according to our data, in a series of rocks and meteorites

Name of material	Absolute ratio of C^{12}/C^{13}	Deviation from the standard in %
Average of three determinations of iron-stone meteorites (Bragin, Lipetskiy farmstead, Pallasite)	91,54	—2,56
Same	91,60	—2,62
Canyon Diablo meteorite	91,55	—2,57
Same	91,53	—2,55
Dunite, North Siberia Sob'-Voynar, Sample 1023)	91,64	—2,66
Gabbric diabase	91,61	—2,63
Western Siberia (Sample 1028)	91,66	—2,69
Kirgiz granite (Sample 205)	91,17	—2,15
Same	91,18	—2,17
Same	91,14	—2,13

NOTE: Comma represents decimal point.

In summing up the data on isotopic ratios of sulfur, oxygen, and carbon, it may be said that these ratios are very similar for all types of meteorites. The new high temperature phases, such as those of Fe_3C, provide an exception. The above ratios for different igneous rocks have a broader range, and they differ from the isotopic ratios in meteorites insofar as the heavy isotopes O^{18}, S^{34}, and C^{13} tend to increase in acid rocks. In isotopic ratios of light elements, the dunites, more than other rocks, resemble meteorites (chondrites).

All this allows us to state that the formation of meteorites occurred in regenerative circumstances, at higher temperatures than those prevailing in the process of magmatic differentiation on earth, and that the separation processes which occurred during the formation of meteorites differ from processes occurring in the separation of magma.

5. ZONAL FUSION OF CHONDRITES AND SILICATES

The character of the distribution of elements, in particular of rare elements and isotopes in chondrites and achondrites, as well as repeatedly stated considerations as to the origin of achondrites from chondrites, pointed the way directly to experiment. Simple fusing-out from chondrites of the more fusible phase, as our experiment indicates, does not bring about a division of the silicates of the chondrites into distinctly separate phases. For this reason we turned to zonal fusion.

When the substance of a meteorite, in the shape of a thin rod, is heated in a narrow zone until fusion is reached, and when the heater is then slowly moved along the rod in such a way that the meniscus of the liquid phase continually passes above the solid phase, and when this procedure is repeated many times, then the substances which lower the fusion temperature pass over to the liquid phase, whereas the substances which increase the temperature remain in the solid phase. This is the so-called zonal fusion process. The smaller the distribution coefficient $K = \frac{K \text{ solid phase}}{K \text{ liquid phase}}$, the greater the efficiency of the separation. The value of this process consists in the fact that the liquid phase repeatedly moves along a considerable distance above the solid phase, thus becoming enriched with fusible (and readily volatile) components. In this process, the movement is in one direction, with the entire liquid phase being pushed aside. It seems to us that this separation process is not merely an image of the fusing-out of the earth's crust on our planet.

Zonal fusion proceeds very well when we have a one-phase system. But the process is not so smooth in binary and more complex systems; owing to partial disintegration of the system, and because of the formation of eutectics and aggregates, the process of zonal fusion is somewhat limited. The process becomes even more complicated in the presence of volatile components.

We carried out zonal fusion with a non-ıagnetic fraction, that is, with the silicate hase of a meteorite — the Saratove chondrite which fell in 1918). It is an unaltered chon-rite consisting of rather large chondrules. 'hey are as much as 5 millimeters in diam-ter. The granular mass of the meteorite is emented by fragments and dust of its own hondrules. Under the microscope one can ee in thin sections, irregularly fused chon-rules; they consist of olivine or of olivine nd rhombic pyroxene. The meteorite con-ains approximately 7.25% iron and 6.52% 'eS.

An analysis of the Saratov meteorite was arried out by L.S. Selivanov, separately rom the magnetic and non-magnetic frac-ons. By repeated demagnitization, approxi-ıately 11% magnetic and 89% non-magnetic actions were obtained.

The following is the composition of the on-magnetic fraction of the Saratov meteor-e (in %):

SiO_2—44.83	MgO—27.43
TiO_2— 0.12	MnO— 0.39
Al_2O_3— 1.80	Na_2O— 0.56
Cr_2O_3— 0.71	K_2O— 0.10
FeO—14.46	P_2O_5— 0.74
CaO— 2.33	

Without going into details of chemical omposition for the average chondrules, it ıay be said that they are formed of two ıain molecules: olivine Mg_2SiO_4 (containing e_2SiO_4~25%) and rhombic pyroxene $MgSiO_3$ containing $FeSiO_3$~25%). The phase diagram f the MgO—SiO_2 system, which gives the $MgOSiO_2$ and $MgOSiO_2$ phases, has been ufficiently investigated. We deem it impor-ınt that in this system $2MgOSiO_2$ fuses at a ery high temperature, approximately ,800°C, whereas the lowest temperature f the liquid phase of this system corresponds an $MgOSiO_2$ composition that melts incon-ruently according to the formula:

$$2MgSiO_3 = Mg_2SiO_4 + SiO_2.$$

Apparently, all other components of the ilicate phase in chondrites form part of the hases Mg_2SiO_4 and SiO_2. Investigations ere carried out to determine more or less xactly the content of olivine and pyroxene chondrites (Table 20). If we accept Wahl's ata, we can calculate the quantity of SiO_2 hich is obtained upon fusion of the chondrite.

Pyroxene content (%)		
$CaSiO_3$— 2.26	0.58	SiO_2
$MnSiO_3$— 0.51	0.14	"
$MgSiO_3$—16.50	4.93	"
$FeSiO_3$— 7.36	1.67	"
26.63	7.3	

Consequently, 7.3% SiO_2 is contained throughout the chondrite. The olivine fraction consists of $MgSiO_4$ 30.67 + Fe_2SiO_4 13.85 = 44.52 throughout the chondrite. Thus, at a temperature of 1,557°C, it is possible to fuse out a maximum of 7.3% SiO_2 (plus various admixtures).

From a number of phase diagrams of Fe and Ca oxides, and a number of rare elements with SiO_2, such as alkaliés, alkaline earth elements, and others, it follows that all these aggregates — admixtures in the MgO—SiO_2 system — will lower the fusion temperature of $MgSiO_3$. Also, many metallic oxides will wholly or partially form aggregates with SiO_2 (which is unimportant, because they are found in very small quantities in chondrites), and will immediately dissolve with SiO_2 to form metasilicate salts, such as those of Fe, U, Th, and others, which decrease the fusion temperature. On the other hand, many metallic oxides are stable under these con-ditions, in the form of orthosilicates, as for example Ni, Co, partially Fe, Zn, Cd, and others. This means that they will give solid solutions with Mg_2SiO_4, which decompose, as a rule, at considerably higher temperatures than the fusion temperature of metasilicates. Finally, certain oxides do not form solid solutions with either SiO_2 or Mg_2SiO_4. Thus, for example, Cr and others, or also olivine metasilicate, decompose at temperatures higher than 800°C, etc. Consequently, at the incongruent temperature of fusion of $MgSiO_3$ from chondrites, one can expect the fusing-out of approximately 7% SiO_2. This is a very important condition which limits our entire discussion of the fusing-out mechanism.

The zonal fusion of the chondrites' silicate fraction was effected in the following way. A fine powder (approximately 100 mesh) of the silicate fraction of the chondrites was pressed in the form of a column, usually 12 to 18 centimeters in length and one centi-meter in diameter. This required about 50 grams of the non-magnetic fraction.

After many tests concerning the best material and the best form of the boat-shaped device, as well as other experimental con-ditions, it was found most convenient to use this device in the form of a small ribbed tube made of pure graphite, about 25 centimeters in length, with an outside diameter of about 3 centimeters and an inside diameter of 1.2 centimeters. During the experiment, this tube of graphite was placed in a quartz tube through which currents of argon passed. Heating was done from the circuit of a high frequency furnace; the fusion zone was 0.5 centimeters wide; the speed of motion of the heater was 0.5 centimeters per hour. After preliminary sintering of the silicate mass of the chondrites at 1,200°C, zonal fusion was

Table 20

Zonal fusion of the Saratov Meteorite

Samples	SiO_2	Al_2O_3	TiO_2	MgO	CaO	K_2O	Na_2O	FeO
Fusion 1:								
0—1 millimeters	65,30	1,12	—	16,6	1,78	—	—	—
1—4 "	58,9	3,63	—	32,3	3,02	—	—	—
4—6 "	51,3	5,02	—	27,5	2,51	—	—	—
85—87 "	44,2	2,14	—	25,7	2,0	—	—	—
122 "	47,9	2,29	—	25,7	1,88	—	—	—
160 "	45,8	—	—	26,3	2,4	—	—	—
Fusion 3:								
0—1 millimeters	52,6	1,82	0,14	—	2,29	—	—	—
1—2 "	49,0	2,40	0,11	28,2	2,35	—	—	—
2—3 "	42,7	1,95	0,1	28,2	2,14	—	—	—
8—10 "	38,9	1,25	0,1	27,5	3,47	—	—	—
13—14 "	44,7	1,48	0,11	24,3	2,19	—	—	—
17—15 "	44,9	1,85	0,11	28,2	2,41	—	—	—
Composition of the Saratov chondrite (incomplete fraction)	44,83	1,80	0,12	27,43	2,33	—	0,56	14,46
Average composition of chondrites (silicate phase)	47,0	3,09	0,14	29,48	2,41	0,21	1,2	15,40
Average composition of achondrites (non-feldspathic)	52,56	1,09	0,12	30,47	1,20	0,11	0,36	11,45

NOTE: Comma represents decimal point.

effected, usually at 1,500 ± 50°C. Between 1,800 and 2,000°C, the entire chondrite was sublimated. Five to seven passages were made along the entire length of the column of the silicate mass of the chondrites, which took more than 40 hours. Results are given below.

The apex of the column of the chondrite substance, to where the fusible phase was driven, changed both in structure and in chemical content. Naturally, the chondrules disappeared. Thin sections show that at the apex of the column a vitreous mass with occasional crystals of rhombic pyroxene developed (Fig. 5).

The quantity of crystals increases in a thin section, cut 4 millimeters from the apex (Fig. 6). A completely different situation exists in the middle part of the column, for example at a distance of 85 millimeters from the apex. Here, well-formed crystals of rhombic pyroxene, as well as notable amounts of olivine, appear (Fig. 7). They reach considerable size in the shape of elongated crystals, in thin sections cut from the end of the column opposite its apex (at a

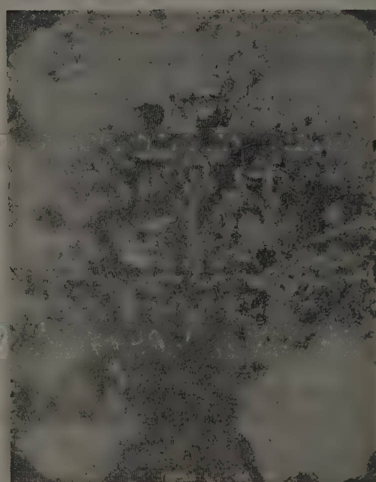

FIGURE 5. Thin section from a silicate phase after zonal fusion, at a distance of 1 mm from the apex. Magnification 75X.

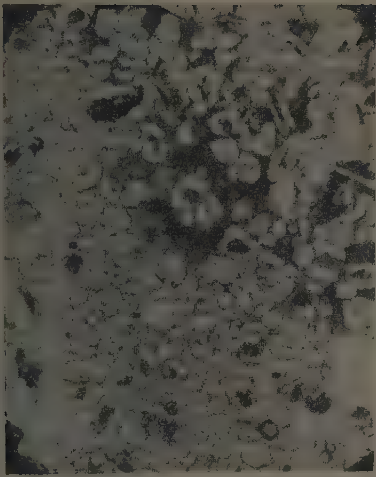

FIGURE 6. The same, at a distance of 4 mm from the apex. Magnification 75X.

listance of about 160 millimeters), where here is also much olivine (Fig. 8).

Chemical analysis has shown (Table 20) hat the SiO_2 was most mobile; its concentration increased considerably at the apex of the column. There was very little movement of Al_2O_3, CaO, and TiO_2. The K_2O content hardly changed along the entire length of the column. The content in fusions 1 and 3 from 0 to 6 millimeters (see Table 20) can be compared with the average content of non-feldspathic achondrites.

According to calculations, approximately 2% of the SiO_2 in the silicate fraction are pressed out. When zonal fusion is applied to the silicate fraction of the chondrites, there is only a very small amount of volatile aggregates. Only sulfur is sublimated and condensed on the walls of the quartz tube. This is connected with the very nature of the fusing-out method.

It was highly instructive to compare the behavior of similar silicates in the presence of perceptible amounts of volatile aggregates, for example of F. A mixture was taken, essentially similar to MgO 37.5:SiO_2 62.5.

SiO_2—59.0	CaFe—0.75
MgO—35.5	TiO_2—0.40
Al_2O_3— 2.30	ThO_2—0.1
CaO— 0.95	UO_2—0.1
KF— 0.90	La_2O_3—0.1

By means of sintering MgO and SiO_2, the metasilicate $MgSiO_3$ was obtained. This metasilicate was posdered and all the remaining components were added to it. Zonal fusion was applied in the same manner as for meteorites. Before the incongruent melting point of $MgSiO_3$ was reached, few sublimates were formed on the walls of the quartz

FIGURE 7. The same, at a distance of 85 mm from the apex. Magnification 75X.

FIGURE 8. The same, at a distance of 100 mm from the apex. Magnification 75X.

Table 21

Zonal fusion of silicate mixture 1

Samples	SiO_2	TiO_2	Al_2O_3	MgO	CaO	K_2O
Fusion 4						
1. Encrustation	67,5		1,07	20,09	1,10	13,2
2. Light fraction, 10 cm	63,1	0,56	2,51	33,2	2,75	—
Same, 3 cm	62,5	0,52	2,10	33,9	2,63	0,42
Same, 11 cm	60,3	0,16	2,09	26,9	1,90	—
3. Heavy fraction, 13 cm	53,7	0,075	3,58	40,6	3,10	—
Same, 12 cm	59,0	0,58	2,58	35,5	2,85	0,34
Same, 7 cm	55,6	0,51	2,63	33,8	2,44	—
4. Composition of the silicate mixtures	59,0	0,40	2,30	35,5	1,7	0,72

NOTE: Comma represents decimal point.

and graphite tubes. At the moment of incongruent fusion, approximately 1,600°C, when the crystalline lattice structure of the silicate abruptly broke down, there was a considerable amount of volatile sublimates. According to analysis, these sublimates contained the main mass of potassium; later, uranium and other elements were sublimated (Table 21).

We observed a sharp change in the mobility of a number of elements. The fusion presented a very complicated system. In the silicate column there were masses with higher or lower quantities of SiO_2 and other elements.

Thus we observed a difference between the "dry" process of zonal fusion and the process with volatile aggregates.

6. DISCUSSION AND CONCLUSIONS

If we accept, as do most authors, that the mantle of the earth was composed of a substance similar in content to chondrites, and furthermore that the earth's crust was formed by the fusing-out and degassing (at the expense of the radioactive heat[2] of U, Th, and K^{40}) of the mantle in a way analogous to zonal fusion, vertically, along the earth's radii, then we may estimate the thickness and characteristics of the earth's crust and of the crust of other bodies. In this process of crust formation by fusing-out and degassing, the length of the path of molecular diffusion should be determined, because evidently the possible thickness of the crust will depend on the radius of the planet. The earth's crust is very thin, and according to various authors its thickness is 30 to 70 kilometers, i.e., 1 or 2% of the thickness of the mantle. Zonal fusion suggests, as do calculations show that the maximum possible thickness of a crust may reach about 7% of the thickness of the mantle. We think that residual rocks from the fusing-out of the mantle substance are dunites. These rocks are known to emerge sometimes on the earth's surface in the form of elongated masses, which fill strips of deep faults of the earth's crust.[3] These are the rocks of the mantle. As a rule, dunites contain nearly 2% pyroxene. Therefore, if the mantle were completely fused, a 6% fusible silicate yield could result. This explains, it seems to us, the thinness of the earth's crust.

One may estimate, though only very roughly, the thickness of the mantle which determines the concentration of the various chemical elements in the earth's crust.

[2] Probably, at the moment of maximum compression of the substance of the proto-earth, gravitational heat appeared, only to decrease rapidly. The heat of the radioactive disintegration of U^{235}, U^{238}, Th^{232}, K^{40}, and other elements began by that time to increase only gradually, reaching its maximum at a later stage in the evolution of the earth.

[3] The occurrence of Fe, Ni, Co, Pt, Cr, and some other elements is connected with these and certain other olivine rocks in the earth's crust. The way in which they were formed depended, probably, on the action of rocks in the saturation zone (of the earth's crust) upon the most external part of the mantle rocks which happened to be within the saturation zone of the crust.

Table 22

Mantle thickness, necessary for the concentration of various elements

Elements	Silicate phase in chondrites in %	Dunites, in %	Basalts, in %	Granites, in %	Thickness of mantle in kilometers
Li	$3\cdot10^{-4}$	$1\cdot00^{-4}$	$1{,}5\cdot10^{-3}$	$7\cdot10^{-3}$	~400
K	$8{,}5\cdot10^{-2}$	$2{,}5\cdot10^{-2}$	$8{,}3\cdot10^{-1}$	3,34	~700
Rb	$8\cdot10^{-4}$	$2\cdot10^{-4}$	$1\cdot10^{-2}$	$4\cdot10^{-2}$	~850
Cs	$6\cdot10^{-5}$	$(5\cdot10^{-6})$	$(2\cdot10^{-4})$	$1\cdot10^{-3}$	~250
F	$(1\cdot10^{-2})$	$1\cdot10^{-2}$	$4\cdot10^{-2}$	$8\cdot10^{-2}$	~100
Cl	$1\cdot10^{-2}$	$4\cdot10^{-3}$	$2\cdot10^{-2}$	$2{,}5\cdot10^{-2}$	~100
Br	$3\cdot10^{-3}$	$1\cdot10^{-4}$	$4{,}5\cdot10^{-4}$	$2\cdot10^{-4}$	—
Al	1,74	0,44	8,76	7,70	~150
Ga	$5\cdot10^{-4}$	$2\cdot10^{-4}$	$2\cdot10^{-3}$	$3\cdot10^{-3}$	~150
Ti	1.10^{-1}	$6\cdot10^{-2}$	$9\cdot10^{-1}$	$2{,}3\cdot10^{-1}$	~400
Ca	1,97	0,5	6,72	1,58	~100
Sr	$1\cdot10^{-3}$	$2\cdot10^{-4}$	$4\cdot10^{-2}$	$3\cdot10^{-2}$	~100
Ba	$8\cdot10^{-4}$	$6\cdot10^{-4}$	$2{,}7\cdot10^{-2}$	$4{,}5\cdot10^{-2}$	~3000
U	$2\cdot10^{-6}$	$1\cdot10^{-7}$	$8\cdot10^{-5}$	$3{,}5\cdot10^{-4}$	~1000
Th	$1{,}0\cdot10^{-5}$	$(5{,}0\cdot10^{-7})$	$3\cdot10^{-4}$	$1{,}8\cdot10^{-3}$	~2000
Mo	$5\cdot10^{-5}$	$4\cdot10^{-5}$	$1{,}4\cdot10^{-4}$	$1{,}9\cdot10^{-4}$	~400

Let's assume that the average thickness f granite and basalt over the entire earth urface, is 10 kilometers for each, and that he density of both rocks is the same. Calcu-ations show that the entire thickness of the nantle rocks participate probably in the rocess of fusing-out and degassing. Of ourse, similar assumptions as to the thick-ess of such layers for each of the elements vould entail a considerable margin of error.

It would not be sensible to dwell upon the inetics of fusible and volatile substances arried up to the earth's surface, which esult in a high gradient of concentration of number of chemical elements, and their aturating the external envelope of the earth, hat is, the crust. Later on, fractional differ-ntiation, as well as other accompanying rocesses of division in this saturation zone f fusible and volatile aggregates, resulted in he tremendous variety in rock composition nd many different phases. Also, we should ote that the entire mass of the volatile ggregates was not concentrated in the crust; n part, the volatile aggregates passed through his screen; only the remainder was left in he crust, for example S, F, B (in skarns) nd CO_2. But the main mass of N_2, H_2O, O_2, HCl, and probably the main mass of nert gases, during the degassing process, vere left outside the igneous rocks and the aturation zone of fusible and volatile aggre-ates. The distribution of these aggregates hroughout the earth's crust differs and does ot correspond to the initial ratios in chon-rites. Processes of degassing and fusing-out rom the mantle are, to a certain extent, oing on even at present; and separation of substances within the earth's crust are doubtlessly still following their course. Furthermore, the isotopic content not only of lead ore but of lead impurities in any other ore — iron, copper, etc. — shows that the process of their breaking away from magma or rock, occurred quite normally at various times. There is no ore envelope formed once and for all. Accordingly, present-day olivine lavas are of recent geologic age, as determined by radioactive methods (Table 22).

Differentiation of meteoritic material is quite different. Here we may assume the existence of two different processes: the first leads to formation of iron-stony chondrites and iron meteorites, and the other to all the remaining types of stony meteorites, the achondrites.

I shall not dwell on the various hypotheses concerning the formation of chondrites. I assume that their similarity in size, as well as in mineralogic and chemical composition, indicates a particular phase condition at some stage of the evolution of meteoritic matter. This condition must have been one of mist droplets from liquid silicate, with resiliency (such as that of steam) sufficient to prevent the drops from merging, and which underwent self-purification by means of repeated recondensation. These drops could be formed, as may be proved experimentally, from the tiniest silicate crystals e.g., olivine, enstatite, and others) when they went through the 1,800°C zone.

According to this point of view the iron-stony meteorites, for example pallasites, are

not actually meteorites resulting from an incomplete process fusing-out iron from the parent meteoritic substance (?) (in which case all the chondrules would have undoubtedly disappeared); quite to the contrary, they are meteorites which formed from silicate droplets and Fe by means of agglomeration. The fact that the vast majority of meteorites consist of separate, commonly deformed chondrules or of small fragments of mineral grains, or finally, of large fragments forming breccia-like meteorites, suggests the great importance of collisions in the process of meteorite formation. These collisions resulted either in the mechanical destruction of the meteoritic material, or in the agglomeration of bodies of various size. Possibly, the olivine of iron-stony meteorites (which is "purest" in content) suggests a maximum temperature of formation for such meteorites. The origin of achondrite is due to a different process.

The chemical composition of all the achondrites, particularly in content of rare elements and isotopic content of light elements (if one derives them from material similar to chondrites), indicates that this process of division was not proceeding in depth; in other words, it took place in bodies of small volume, where the formation of a saturation zone or crust, similar to the earth's crust, could not occur. Whereas the formation of the earth's crust proceeded in the presence of considerable masses of volatile aggregates, the formation of achondrites, for example non-feldspathic achondrites, took place in their complete or almost complete absence (because of the small radius of the body) and at temperatures higher than those assumed for the formation of the earth's crust. Whereas, one can say, that in the first case formation went on in a close isolated system, in the second (that of the non-feldspathic achondrites) the system was more open, resulting in the loss of volatiles and in the dominance of the "dry" process.

The formation of feldspathic achondrites is reminiscent of the mechanism which results in the transformation of the Dinas bricks used as a lining for Martin (open hearth) furnaces, at high temperatures over long periods. Zones of various composition are formed in the Dinas brick; in the transitional zone, Al_2O_3, CaO, and TiO_2 accumulate; and this is not unlike what happens in feldspathic achondrites.

Summing up, I can say that the differentiation of silicate material of large and small bodies depends upon the MgO-SiO_2 system and the distribution of various elements between the Mg_2SiO_4, $MgSiO_3$, SiO_2 phases, in circumstances that vary widely.

Formation of zones of sufficient thickness, with high gradients of concentration of fusible and volatile aggregates, similar to the earth's crust, depends not only on the composition of the parent material, but also on the thickness of the mantle, i.e., on the radius of the planet or other body. As a result of the fusing-out and degassing the residual rock is, under these conditions, dunite. A comparative study of the distribution of a number of elements and isotopic ratios, as well as the method of zonal fusion applied to meteorites and silicates, suggests a difference in character between the process by which meteorites evolved into their several varieties and the processes which occurred on the earth's crust: fractional crystallization, hydrothermal activity, etc. The process (Table 22) goes on in regenerative conditions.

Two fundamental processes are distinguishable in the formation of meteorites: 1) mixing, agglomeration of silicate chondrules and iron masses as well as of troilite, in the formation of chondrites, iron-stony, and stony meteorites; and 2) formation of achondrites from chondrites on bodies of small size, with the result that volatile aggregates are not present in sufficient amounts, or are actually lost.

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V.I. Vernadskiy Institute
of Geochemistry and Analytical Chemistry
of the Academy of Sciences
of the U.S.S.R.
Moscow

Received January 10, 1959

THE ORIGIN OF THE KAZAN-SERGIYEVSK BASIN[1]

by

A. S. Novikova

According to the author, the Kazan-Sergiyevsk basin was formed in a region of pre-Devonian gabbric diabase. It contains Devonian volcanic and sedimentary volcanic accumulations.

* * * * *

Devonian volcanics are found on the Russian platform at Timan and along the southern districts of the Donbass. Recently Devonian extrusive and sedimentary extrusive formations were found in exploratory drillings in the Chernigov region, at Soligalich and also in many places in the Volga-Ural Oblast'; at the settlements of Kazaklar, Radayevka, Gusikha and elsewhere. Lava and tuff sequences, heretofore unknown, are even more extensive. We have already indicated that at the beginning of the formation of the sedimentary mantle on the Russian platform, graben-like basins developed, dividing the crystalline foundation into several blocks, with volcanic layers forming on the borders of the graben-like basins, especially at marginal parts of the platform [13]. The present paper attempts to clarify the tectonic position of the Devonian andesitic and basaltic lavas and tuffs in the eastern part of the Russian platform.

In pre-Devonian time the Volga-Ural crystalline massif was a relatively elevated region. Reef sediments are either absent or of insignificant thickness. The granites associated with this group are hidden by more recent formations. On the basis of drilling and geophysical investigations, the eastern and especially southeastern areas of the uplift are angular in outline; the marginal parts of the massif are marked by depressions filled with pre-Devonian sediments; the Radayevka depression, the Bavlinsk and Zol'nensk Ovrag depression in the Samara bend of the Volga region. In the Bavlinsk depression, the total thickness of the pre-Devonian sediments reaches 540 meters, but in the Buguruslan region drillings penetrated to 600 meters without reaching their base. In the adjacent uplifted sections, as in the remainder of the massif, pre-Devonian formations are absent and Devonian sediments lie directly on crystalline rocks. A special feature of the structure of the pre-Devonian Volga-Ural crystalline massif is the gabbric diabase series. It is mainly concentrated in its eastern half along the edge of the Tatar uplift.

In Middle Devonian times sedimentation began on the Volga-Ural massif. On the western slope, the Devonian sediments are similar in composition and fossil fauna to the formations in the central areas of the Russian platform. On the Tatar projection of the basement, terrigenous sediments of insignificant thickness accumulated. In the western marginal section of the Tatar uplift and in the Ul'yanovsk-Chuvash projection Middle Devonian Pashiysk and Kynovsk sediments are either missing or are of an insignificant thickness. The Kazan-Sergiyevsk basin was formed on the arched areas of the Volga-Ural elevation, where the pre-Devonian volcanic formation is present. This basin is outlined by a relatively complete section of Devonian terrigenous formations of considerable depth, distinguished by their narrow, elongated form.

The map shows the sedimentary structure of the Volga-Ural crystalline massif at the end of Kynovsk time; stratigraphic columnar sections of its components are given in the table.

1. Sedimentation in the Kazan'—Sergiyevsk Basin and Adjacent Structures

Of the sediments filling the basin, those of interest occur between the base of the Devonian section and the Kynovsk strata of the lower Frasnian. This part of the section can be divided according to lithology and index

[1] O proiskhozhdenii Kazansko-sergiyevskogo progiba.

fossils into Middle Devonian and Lower Frasnian. Middle Devonian formations occur at the base, the Eifel formation (Takata beds) and the lower and upper Givetian beds; the upper coincides with the Lower Frasnian substage of the Upper Devonian in the Pashiysk and Kynovsk sequences.

The Takata beds of the Eifelian are developed in the southeastern part of the basin near the village of Sterlibashevo. M.F. Mikryukov [8] indicated that they overlie the eroded pre-Devonian formations. The lower part of the section consists of gray and brownish-gray sandstone and weakly sorted gravels attaining a thickness of 25 meters; above this lie sandstone and siltstone nearly 12 meters thick. Highly ferruginous shales containing oolites of brown iron ore and chamosite are interbedded in the upper and lower surfaces of the Takata layers. The thickness of the Takata beds decreases from 10 to 3 meters northwest and west of Sterlibashevo. They are also found in the sections at Buguruslan and at Baytugan, Bavly and other villages. No Eifelian sediments are present outside the basin.

More complete sections of the Lower Givetian and Eifelian are found along the southern borders of the basin from Sterlibashevo to Buguruslan. In the Sterlibashevo region, the Takata sandstones are overlain by gray quartzite, arenaceous to argillaceous conglomerate with limestone fragments. Their total thickness is 1.5 meters. At the bottom ferruginous-chamositic oolites are present in clay interlayers. Above lie limestone and black bituminous clay with fragments of Favosites ex gr. goldfussi Orb., Alveolites sp., Conchidium baschkirikum (Vern.), Pentamerella arata (Conr.), Uncynulus goldfussi Schnur., Atrypa ex. gr. aspera Schl., Dechenella sp.

The total thickness of the Lower Givetian sediments in Sterlibashevo region reaches 50 meters [8]. In the Buguruslan region, the Lower Givetian formations begin with gray argillaceous carbonate rocks with interlayers of coarse-grained kaolinized sandstone (thickness 15 meters). Above this there are approximately 20 meters of finely crystalline dolomite alternating with argillaceous limestone with many brachiopods, pelecypods, tentaculites and crinoids.

In the central parts of the transverse belt, in the region of Sernovodsk, the lower Givetian formations are represented by limestone with corals and ostracods. The outstanding feature of the northern part of the basin is the presence of ferruginous-chamositic oolites. Oolitic ores of considerable thickness are found in the Tuymazy region, where the lowest Givetian formations, according to data by M.F. Mikryukov and K.R. Timergazin [7], mainly consist of sandstone 5 to 10 meters thick, overlain by argillaceous-calcareous formations 5 to 20 meters thick. The lower sandstones are bright yellowish gray, cemented by argillaceous-kaolinitic material. Higher, in the interlayers of clayey siltstone, clay and ferruginous-chamositic oolitic ores appear. The clay consists of kaolinite containing more than 20% Al_2O_3 [3]. The ores consist of oolites of chamosite and brown ferric hydroxides, 0.6 to 1.2 millimeters in size. They form lenses and layers 0.5 meter to several meters thick in the lower and upper parts. The rocks contain up to 60% oolites. They are unevenly distributed in a kaolinitic material containing also alumo-chamosite, in places calcite and rarely pyrite. The structure of the oolites is concentric, laminar and radial, and are usually oval and occasionally spherical [10]. Above are argillaceous carbonate rocks, with interlayers of bituminous pteropodic limestones of the Domanik type.

The amount of carbonate rocks markedly decreases towards the northern flank of the basin and they are replaced by arenaceous and argillaceous formations. In the uplifted parts of the basement, north of the villages Bakaly, Aksubayevo and in the region of Melekes and Uzyukovo villages no Lower Givetian foundations can be found. They reappear near the villages of Kazaklar and near Sovetsk, along the western border of the Tatar elevation. Near Kazaklar, the granitic basement is overlain by nearly 2 meters of pyritic limestone and sandy clay with abundant remnants of Aparchites sp. Near Sovetsk the crystalline rocks are overlain by dark-gray argillaceous limestone and clays with remnants of Lingula bicarinata Kut., rare tentaculites, members of the Cupressokrinus species and ostracods. These sediments are 13 meters thick [16].

Thus, in Lower Givetian times the Kazan-Sergiyevsk basin did not possess the features which it acquired later — at the end of Kynovsk time. In certain of its southeastern parts, sedimentation had already occurred in pre-Devonian time, while in the Lower Givetian the area of sedimentation increased considerably, stretching towards the submerged southeastern borders of the Volga-Ural massif forming the Sergiyevsk depression. During Lower Givetian time the basin in the region of the Zainsk and Golyushurma villages branched off from the Sergiyevsk syncline east of the western Tatar uplift and also in the Bakaly region. During the same period the basin in the region of Sovetsk and Vozhgalya, began to sink separating the rest of the massif from the west side of the Tatar uplift.

This structure was isolated from the Sergiyevsk depression by the relatively uplifted land

between Kazaklar and the Kamskoye Ust'ye village.

Upper Givetian sediments covered the Lower Givetian depression near Sovetsk and Sergiyevsk, and also the uplifted lands between them, in the Kazaklar-Kamskoye Ust'ye region; another branch developed from the Sergiyevsk depression and from the Kazan-Sergiyevsk basin. We shall now consider the Upper Givetian sediments filling this structure.

In the Sterlibashevo region their structure is as follows:

1. On conglomerate-like limestone with phosphoritic pebbles there are arenaceous and argillaceous rocks containing many fragments of fish bones, brachiopod shells and ostracods; their thickness does not exceed several meters.

2. Above this layer, appear crystalline or argillaceous limestones which are dolomitized, in places bituminous, with remnants of pteropods, Tentaculites and Styliolina, and also brachiopods, crinoids and large ostracods; (thickness 35 meters).

3. Above lie clay and sandstone with siderite concretions, plant detritus and lingula shells; (thickness 10 to 12 meters).

4. The section ends with argillaceous limestone and bituminous clay with remnants of brachiopods, crinoid segments, coral fragments and stromatoporoids; (thickness 10 meters).

According to V.N. Tikhiy [16] the Upper Givetian in the central part of the Sergiyevsk basin in the Baytugan-Sernovodsk region, forms the following sequence:

1. Sandstone, gravel and argillaceous siltstone interbedded with carbonaceous shale. The cement is argillaceous, ferruginous and in places basaltic, carbonate and kaolinitic. Higher there appears a rubble of argillaceous shale, fragments of ostracod and pelecypod shells, and spores characteristic of Mosolova layers of the central region of the platform. (Total thickness 12 to 19 meters).

2. Alternation of clastic argillaceous organic limestone, fine-grained dolomite, and bituminous clay with abundant remnants of Alveolites suborbicularis Lam., Ilmenia subumbonata Hall., Spirifer cf. pseudopachirinchus Tsch. Higher there are clays with plants: Archaeozonotriletes extensus Naum., A. vulgatus Naum. and A. pustulatus Naum. (Thickness 12 to 16 meters).

3. Bright gray and white sandstone with argillaceous, sideritic, and calcitic cement, 30 meters thick; argillaceous siltstone interbedded with gray clay, with siliceous and sideritic concretions with plant remnants and spores, 10 meters thick; gray and black-brown sandstone, siltstone and clay, thin-bedded, plicated, laminated, interbedded with argillaceous siderite and chamositic oolites. (Total thickness 45 to 66 meters).

4. Dark gray, almost black limestone with layers of organic-fragmental limestone containing Liturella dubia (Mit. et Carm.), L. fragillis Byk. (in lit), Lingula bicarinata Kut. Above this layer the carbonates are replaced by gray clay with siderite concretions containing remnants of Ilmenia subumbonata Hall. and Tentaculites Luasch. This is terminated by dolomitic limestone; on the surface of this layer, remnants, of Buchiolia misera Holz. and small Lingula were preserved. (Total thickness is 17 to 22 meters).

5. Gray, thin-bedded siltstone, similar to that of the upper part of 3; 20 to 30 meters thick.

6. Gray clays with large siderite concretions and remnants of Lingula, fish bones and ostracods, and also slightly recrystallized, in places bituminous limestone, and in places glauconitic. Among the fossils are Syringopora sp., Favosites ex gr. serricornis Blein., Productella ex gr. subaculeata Murch. var mesocevnoica Nal. (?), Atrypa ex gr. aspera Schl. (Thickness is 10 to 17 meters).

We found that the Upper Givetian formations of the central part of the Sergiyevsk depression extend into the Bavly-Tuymazy region; this was determined by the similarity of sediments developed there in the above-described sections. In Tuymazy, the denuded top of the Lower Givetian limestones are often overlain by an unusual rock 1 meter thick, consisting of kaolinitic argillite in places becoming argillaceous conglomerate. It contains up to 36% Al_2O_3 [16]. V.N. Tikhy, in agreement with V.I. Troyepol'skiy, noted that sections in the Bavly-Tuymazy region contain considerably more included and interbedded siderite and pyrite concretions than sediments of the same age which accumulated in the central part of the Sergiyevsk basin.

In this zone, chamosite and ferric hydroxide oolites, and kaolinite and chlorite interlayers are more common. The total thickness of the Upper Givetian formations here does not exceed 65 meters, because of consequent erosion.

Dark-gray argillaceous limestone and clay with lingulae, pteropods, crinoids and ostracods, especially characteristic of the Mosolovian, are conventionally referred to the Upper Givetian in the northern parts of the

Kazan-Sergiyevsk basin, in the Sovetsk region. (Total thickness 43 meters). Higher, there is a homogenous layer of white medium- and fine-grained sandstone interbedded with brownish-gray clay and siltstone. Spores of a Staryy Oskal aspect are found. (Total thickness 238 meters) [4, 16].

Upper Givetian sediments are also fully represented in the central part of the basin, between Sovetsk and Sergiyevsk. In the villages Kamskoye Ust'ye and Kazaklar, arenaceous and argillaceous rocks and carbonate bearing remnants of marine fauna alternate. The total thickness of the Upper Givetian sediments is 75 to 125 meters [16].

The data on composition and thickness of sediments presented above show that in the Upper Givetian, the Sergiyevsk and Sovetsk depressions formed one single syncline.

The distribution of the thickest sections of the Pasha sediments in the basin generally coincides with the distribution of the most complete sections of the Upper Givetian.

Gray quartzitic sandstone with abundant carbonaceous detritus, with interlayers and lenses of greenish-gray argillaceous siltstone and clay in the Sovetsk region are conventionally correlated with the Pashiysk sediments. They are 126 meters thick. In the central parts of the basin the Pashiysk sediments are represented by sandstone alternating with siltstone and argillite with occasional interlayers of limestone containing remnants of brachiopods, ostracods and pelecypods.

In the Sergiyevsk depression, in the Shugurov ragion, the Upper Givetian is overlain by mottled bright brown and gray siltstone, dark greenish-gray argillite with many spherulitic siderite accumulations, and also with lenses and interlayers of ferruginous-chamositic oolitic ores. The middle section consists of limestone, separating the lower, arenaceous part from the upper, more argillaceous part. Above the argillaceous layer there is gray dolomite, 2 to 3 meters thick. The total thickness of the Pashiysk layers in the Shugurov region is 70 meters, and 20 to 80 meters in the Tuymazy region. In this region they contain considerably more siderite. In borehole No. 396 at a depth of 1,840 to 1,845.2 meters, and in No. 404 at 1,651.7 to 1,657.8 meters, ferruginous-chamositic oolitic ores were found among the Pashiysk sediments [11].

The volcanic formation found in the region of the village Kazaklar and near Radayevka are characteristic of the Kazan-Sergiyevsk basin. Near Kazaklar the layers containing characteristic Upper Givetian fossils are overlain by a sequence of alternating sandstone, siltstone and argillite containing plants and a few Lingula sp. These correspond to the Pashiysk formations at the villages Verkhniy Uslon and Kamskoye Ust'ye, containing carbonates with marine fossils having characteristics similar to the lower Frasnian [17]; the terrigenous sediments are 25 meters thick.

Vitreous and partly recrystallized andesitic basaltic lavas with ashy tuff on the top appear higher in the section. The volcanic layer is 45 meters thick. They are overlain by gray organic fragmental limestone with many remnants of Kynovsk fossils. In the village of Radayevka, according to K.I. Lomot' [6], boreholes No. 3 and 4 showed that among argillaceous silty sediments in the upper Pashiysk layers there are dark grayish-green sedimentary volcanic formations. They consist of angular 0.06 to 0.2 millimeter quartz grains and 2.5 millimeter fragments of porphyritic extrusives. Fragments of plagioclase crystals are also found. The clastic material is cemented by a hydromicaceous material, admixed with small chlorite scales. Occasionally calcitic secretions are found. The cement comprises 25 to 40% of the rock.

In the basement projections bordering the Kazan-Sergiyevsk basin, Pashiysk and Givetian sediments are either missing or are of insignificant thickness. They are entirely lacking on the west and east of the meridional belt of the basin, and also on the southern, Pokrova-Buzuluk projection of the basement (Fig. 1).

Kynovsk sediments cover the Tatar uplift with a continuous mantle, and also cover the Kazan-Sergiyevsk basin. Various degrees of completeness and thickness of Kynovsk deposits are present in only these structures. The Kynovsk is more fully represented and thicker in the Kazan-Sergiyevsk basin, i.e., in places where the most complete and thick sections of Pashiysk and Givetian are developed. Thus, according to V.N. Tikhiy [16], the base of the Kynovsk, on the northern slopes of the Sergiyevsk basin, consists of a layer 1 to 3 meters thick of gray and greenish-gray limestone and dolomite, with admixtures of argillaceous-bituminous material with many remnants of Lingula punctata Hall., Schizophoria ivanovi Tschern., Productella sericea Buch., Athyris concentrica Buch., Cyrtospirifer murchisonianus Vern. and many other brachiopods, with remnants of corals, crinoids, pteropods and fish. Higher are greenish-gray and brown clay with alumo-chamositic oolites 0.5 millimeters in size, with siderite interlayers and pyrite deposits. In the middle of this layer there are argillaceous greenish-gray limestones, 1 to 7 meters thick.

The total thickness of this argillaceous layer is 15 to 30 meters.

V.N. Loginova studied the lithology of the Kynovsk rocks and noted that argillite is the dominate rock type of this group; the argillite contains up to 27% Al_2O_3, somewhat less than 1% CaO and has a high K_2O and Na_2O content (up to 5%). The upper part of the Kynovsk layers is composed of limestone interbedded with clay, silt and marl; it is 8 meters thick.

Towards the central part of the Sergiyevsk basin, the thickness of the Kynovsk formations increases to 50 meters, and lower units appear. In the Borovka region, under the limestone corresponding to the lowest layer of the above-described section, there is fine-grained sandstone and clay similar to Pashiysk rocks. Arenaceous and argillaceous rocks, 10 to 15 meters thick, overlie a 2 meter-thick limestone layer containing remnants of ostracods typical of the Kynian: Gravia tuimasensis Pol., G. mustaginovi Pol., Buregia zolnensis Pol., Microcoeloenella optata Pol., Indivisia schigrovskensis Pol., Arcatia aff. longa Zasp. and others; the above-described Kynovsk section is found in the central parts of the Kazan-Sergiyevsk basin, in the Kamskoye Ust'ye region, where its thickness is estimated as 50 meters.

According to Ye.N. Larionova [4] the Kynovsk beds in the northern parts of the basin, near Sovetsk can be divided into three layers: the lower, consisting of greenish-gray clay alternating with siltstone, thin interlayers of argillaceous limestone, and containing Megaphillum paschiense Soshk., Cyrtospirifer ex gr. murchisonianus Vern., Striaproductus sericeus Buch. and Lyropcten cf. ingral Vern.; the middle consists of argillite with small pelecypods of the types Pterochaenia fragilis Hall., Ontario sp. and others; tentaculites and orthoceratides are also present in these sediments; the upper layer consists of argillaceous marl interbedded with limestone containing a mixed fauna and spores characteristic of both the Kynovsk and the overlying Sargayevskiy sediments; the thickness of the three layers is 24, 103 and 70 meters, respectively.

On the Tatar uplift the Kynovsk layers are represented by carbonate and argillaceous rocks parallel to the upper limestone layer of the Kazan-Sergiyevsk basin. They are several meters thick.

Data on the sediments filling the Kazan-Sergiyevsk basin follow. Study of this section shows that accumulations here consist mainly of grayish arenaceous and argillaceous sediments and carbonate rocks with remnants of marine invertebrates. These formations are stratigraphically the most complete sediments in the Volga-Ural shield, as can be seen from the table. Eifelian sediments are actually limited to the synclines studied above. Givetian and Pashiysk sediments are missing from the top of the basement projections, while in the basins dividing the southeastern part of the Tatar uplift into a series of blocks; they are represented by thin arenaceous and argillaceous rocks. The Kynovsk section completely covering the Tatar projections of the basement, is also incomplete.

In the Kazan-Sergiyevsk basin, certain rocks in the complex of terrigenous-carbonate formations deserve attention because of their remarkable composition, despite their insignificant thickness. They contain bauxitic and chamositic iron ores, chamosite-oolitic ores, andesitic-basaltic lavas and tuff, and also bituminous limestone and argillite, and are found on various stratigraphic levels in various parts of the Kazan-Sergiyevsk basin. Chamositic and bauxitic ferruginous clay and marl occur as lenses and interlayers over the entire section from Takata to the Kynovsk included. The alumina content of the bauxitic clays is commonly higher than 20%, and in places reaches 36%. Ferruginous chamositic oolitic ores occur along the northern margin of the Kazan-Sergiyevsk basin in the Tuymazy, Bavly and Shururov regions. They form four layers at the bottom and top of the Lower Givetian sediments, at the lower part of the Pashiysk layers, and in the Kynovsk sediments. The thickness of the ore-bearing layers is 1 to 10 meters, and of the ferruginous oolitic layers, 0.5 to 2 meters. The upper ore-bearing bed contains free alumina [10, 11]. Andesitic-basaltic lava and tuff are developed in the Pasha layers in the vicinity of Kazaklar, Radayevka and also among Upper Givetian rocks in southeastern Tatariya (village of Novo-Yelkhovka). Bituminous limestone and clay are Domanik type sediments; they contain remnants of Buchiolia sp., Tentaculites sp., Pterochaenia fragilis Hall., Styliolina sp. and other. Interlayers of bituminous rocks, sometimes several meters thick, are found in Lower Givetian, Upper Givetian, Pashiysk and Kynovsk formations, predominantly in the marginal parts of the Kazan-Sergiyevsk basin — in Tuymazy, Aksubayevo, Sterlibashevo, and other places [1, 9, 19]. These peculiarities are supplemented by phosphorites, occurring with conglomerate-like limestone in the base of the Upper Givetian sediments in the Sterlibashevo region, and an interlayer of glauconitic limestone in the upper part of Upper Givetian sediments in the center of the Sergiyevsk basin (see table).

Different sediments are found on the western slope of the Volga-Ural massif. Here Middle Devonian and Lower Frasnian sediments, similar to those in the central part of the Russian Platform [20], are developed west of the Ul'yanovsk-Chuvash projection of the basin.

In the Pachelma basin, the Devonian

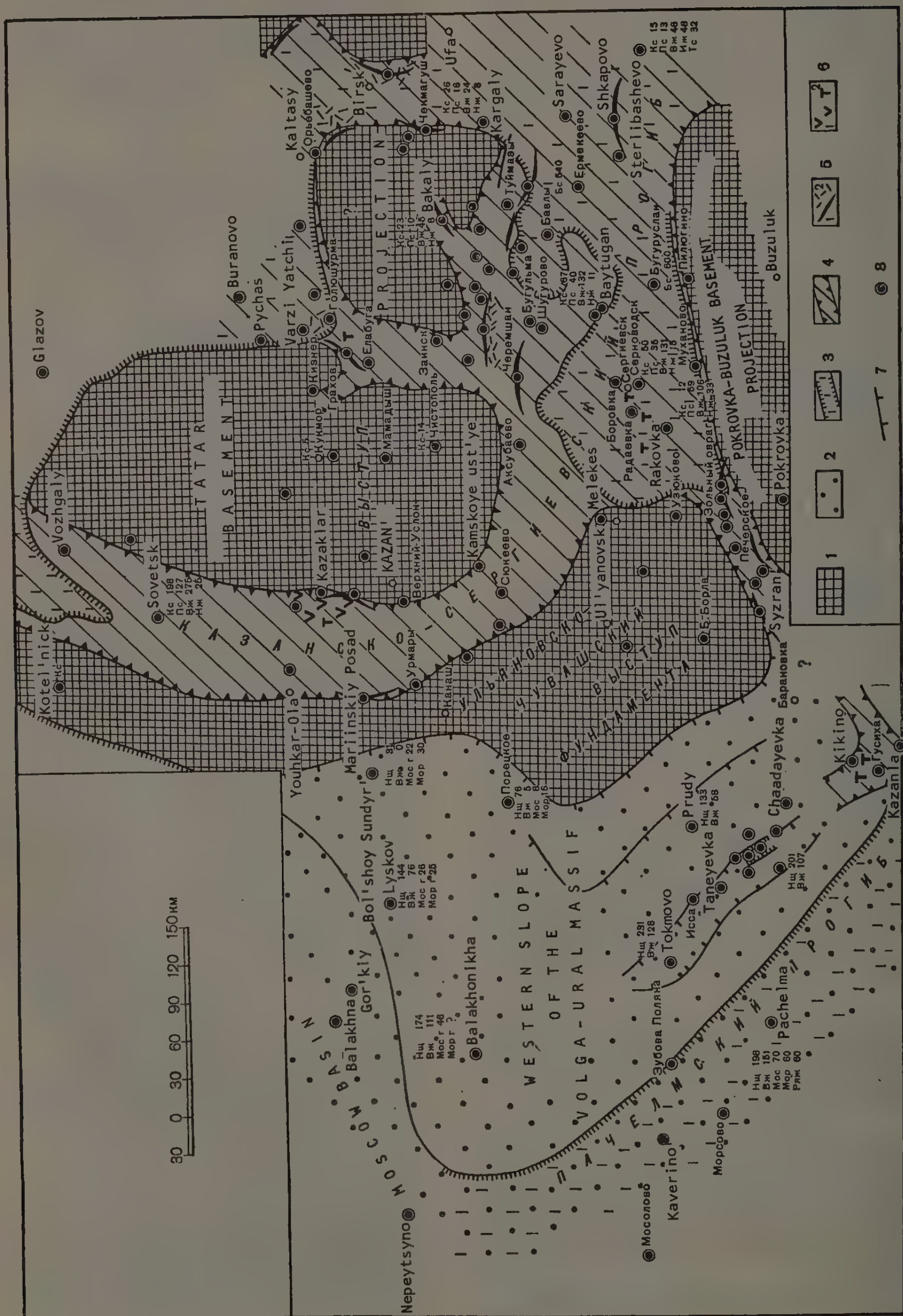
Glazov
Buranovo
Pychas
Varzi Yatchi
Kaltasy
Birsk
Ufa
Kargaly
Bakaly
Sarayevo
Shkapovo
Sterlibashevo
Baytugan
Buzuluk
TATAR BASEMENT
PROJECTION
KAZAN'
Kazaklar
Kamskoye ust'ye
Melekes
Ul'yanovsk
Rakovka
POKROVKA-BUZULUK BASEMENT
PROJECTION
Pokrovka
Syzran
Sovetsk
Vozhgaly
Kotel'nick
Youhkar-Ola
Mariinskiy Posad
Bol'shoy Sundyr'
Lyskov
Balakhonikha
Gor'kiy
Balakhna
MOSCOW BASIN
WESTERN SLOPE OF THE VOLGA-URAL MASSIF
Nepeytsyno
Kaverino
Tokmovo
Taneyevka
Prudy
Chaadayevka
Kikino
Kazanla
Pachelma
30 0 30 60 90 120 150 км
1
2
3
4
5
6
7
8

FIGURE 1. Diagram showing location of volcanic formations in the sedimentary cover of the Volga-Ural crystalline massif, at the end of the Kynovsk.

1 -- Projections of the crystalline basement with reduced sections of Devonian terrigenous formations; 2 -- western slope of the Volga-Ural massif, with Devonian formations filling the synclases in the central part of the Russian platform; 3 -- buried graben-like depressions filled with Devonian terrigenous rocks; 5 -- pre-Devonian hypabyssal rocks (1) and extrusive (2) rocks of the gabbro diabase series; 6 -- Devonian andesite-basalt (1) and tuff (2); 7 -- conventional isopachs on Devonian sediments; 8 -- boreholes.

Map symbols: Бв -- Bavly series; Тс -- Takata layers; [illegible] -- Lower Givetian substage; [illegible] -- Upper Givetian substage; Пс -- Pashiysk series; Кс -- Kynovsk series; [illegible] -- Ryazh unit; Мор -- Morsovo unit; Мос -- Mosolovo unit; [illegible] -- Lower Shchigrovsk unit; [illegible] -- 540, [illegible] -- 33 -- thickness in meters.

formations begin with mottled and gray rocks of the Ryazhsk unit, about 60 meters thick. Higher, sulfate and carbonate rocks of Morsovo layers are about 60 meters thick. No deposition of the Ryazhsk unit occurred on the western slope of the Volga-Ural crystalline massif, and the Morsovo unit is developed only in its northwestern parts, near the villages of Lyskov and Bol'shoy Sundyr', where it lies immediately on the crystalline basement. Here, the lower anhydrite layer is missing and the section is represented by dolomitic limestone 25 to 30 meters thick. Near the village of Poretskoye the Morsovo unit is 16 meters thick. Here the limestone content diminishes, and the clay content rises. These sediments are missing from the rest of the western slope.

The Mosolovo and Morsovo units are developed only in the northwestern part of the Volga-Ural massif. In the Pachelma depression they are 60 to 70 meters thick and are represented by argillaceous and calcareous formations with abundant and varied brachiopods, pelecypods, fish and other organisms. Their thickness diminishes in the eastern side of the depression, and the Mosolovo is completely missing at Tokmovo. The Mosolovo at Balakhonikha in the northwestern part of the massif, is mainly composed of limestone, and in the villages of Lyskov and Bol'shoy Sundyr', of calcareous, argillaceous rocks, 46.26 and 22 meters thick respectively; in the village of Poretskoye it diminishes to 8 meters.

Above the Mosolovo in the Morsovo region there are silty argillaceous rocks with interbedded lenticular siderite accumulations, and abundant remnants of a land flora. The upper part of the terrigenous formations contain much siderite and chamosite. The silty argillaceous Upper Givetian sediments are 151 meters thick. In the southeastern part of the Pachelma basin the argillaceous rocks contain limestone interlayers with brachiopods, pelecypods, gastropods, Lingulidae and other forms, which divide these sediments into the Vorob'yevka and Staryy Osko units. These occur in the western part of the Volga-Ural massif, and towards the central part they pass from calcareous-argillaceous into argillaceous and arenaceous formations. In the terrigenous rocks of the western slope of the massif, as a rule, siderite and chamosite do not appear. The total thickness of Upper Givetian terrigenous formations is 135 meters in the village of Taneyevka, 58 meters in the Prudov region, 5 meters in the village of Poretskoye; they are missing near Ul'yanovsk, the villages of Bol'shaya Borla, Bol'shoy Sundyr' and Mariinskiy Posad.

The lower Shchigry unit corresponds to the Pashiysk and Kynovsk of the Kazan-Sergiyevsk basin. In the Pachelma basin it is represented

Types of cross-sections of the Middle Devonian and bottom of Lower Fransian sediments on the Volga-Ural massif
(according to data of E.N. Larionova, 1956; M.F. Mikryukov, 1955; M.F. Mikryukov and K.R. Timergazin, 1948; N.V. Tikhiy, 1957; M.F. Filippova, 1958; and other investigators)

			Western slope of the massif										Ul'yanovsk-Chuvash basement projection
			Tokmovo		Prudy		Poretskoye						Ul'yanovsk
Upper Devonian	Lower Frasnian subdivision	Lower Shchigry division	Frequently alternating siltstone and sandstone	233 m	Siltstone, sandstone.	133	Silty clay with Lingula rectangularis Ljasch	76	Upper Devonian	Lower Fransian subdivision	Kynian beds		Missing
Upper Devonian	Lower Frasnian subdivision	Lower Shchigry division	Interlayer of sideritic chamositic rocks				White quartzitic sandstones		Upper Devonian	Lower Fransian subdivision	Pasha beds		"
Middle Devonian	Upper Givetian subdivision		Clay, siltstone. Poorly sorted, sideritized sandstone with flora and interlayers of limestone containing Tentaculites, ostracods	128	Sandstone, clay	58	Arenaceous-argillaceous rocks	5	Middle Devonian	Upper Givetian subdivision			"
Middle Devonian	Lower Givetian subdivision	Mosolovo horizon	Missing		Missing		Dark-gray clays	8	Middle Devonian	Lower Givetian subdivision			
Middle Devonian	Lower Givetian subdivision	Morsovo horizon	"		"		Clays, limestones with Lingula bicarinata Kut.	16	Middle Devonian	Lower Givetian subdivision			"
Middle Devonian	Lower Givetian subdivision	Ryazhsk horizon	"		"		Missing		Middle Devonian	Lower Givetian subdivision			
Middle Devonian	Eifelian division		"		"		"		Middle Devonian	Eifelian division	Takata beds		"
Pre-Devonian sedimentary formations			"		"		"		Pre-Devonian formations	Upper and Lower Bavly series			"
Archean													

(continued from preceding page)

Kazan-Sergiyevsk Basin								Tatar basement projection	
Sovetsk		Baytugan-Sernovodsk		Tuymazy		Sterlibashevo		Kukmor	
Marl, limestone rgillites with Pterochaenia fragilis Hall, with Tentaculites, rthoceratites. lays, siltstones.	70 103 24	Limestones, clays	50	Limestones, clays, oolitic chamositic ores	27	Argillites, marls with <u>Cyrtospirifer murchisonianus</u> Vern., siltstones with plant remains	5- 6	Clays, limestones	5
uartzitic sandtones	126	Clays, siltstones, sandstones	35-40	Limestones, clays, sandstones; chamositic ores in the clay layers	20	Limestones with <u>Atripa velikaga</u> Nal., et al. Sandstones, siltstones, interbedded clay and limestones	6-11 10-15	Missing	
Sandstones with ayers of brownish ray clays, siltstones. Clays, argillaeous limestones. vith lingulas, teropods, crinids, and ostraods characteristic of Mosolovo eds	238 43	Alternation of limestones, clays, sandstones, layer of argillaceous siderite with chamositic oolites; glauconitic limestones in upper parts	130	Alternation of limestones, clay, sandstones; layer of limestone, Domanik type clay with lingulas, Crania, Petrocrania, Tentaculites, Goniatites, kaolinitic argillite	[illegible]	Limestone with interlayers of bituminous clays. Clays with siderite concretions. Crystalline limestones with pteropods, brachiopods, crinoids, and ostracods; conglomeratice limestones with phosphorite pebbles.	10 10-12 35	"	
Limestones, lays with <u>Lingula icarinata</u> Kut., vith rare Tentaulites, Cupresocrinus and stracods.	13	Limestones with corals and ostracods	18-31	Limestones interlayered with pteropodic limestones (with Goniatites).	5-7	Limestones, bituminous clays with <u>Favosites</u> e.g. <u>goldfussi</u> Orb., <u>Alveolites</u> sp., <u>Conchidium baschkirikum</u> (Vern.) et al.	50	"	
				Sandstone with layers of clay, containing ferruginous-chamositic oolitic ores; sandstone with argillaceous, kaolinitic cementation	5-10	Conglomeratic sandstones, argillaceous rocks with limestone fragments with interlayers of chamositic clay	37	"	
Missing						Clays with chamositie; sandstones, gravels; ferruginous clays with oolitic chamosite.	37	"	
"		Arenaceous-argillaceous sediments		Dark gray sandstones; arenaceous argillaceous breccia of fragments of gneiss, feldspar, and quartz	140	Arenaceous-argillaceous sediments		"	
Archean								Archean	

by gray and red arenaceous and argillaceous silty rocks, partly sideritized. On the western slope of the Volga-Ural massif, in the upper part of the lower Shchigry argillaceous, marly, calcareous rocks occur in places. The thickest sediments are on the southwestern margin of the massif; in the Tokmovo region, the thickness reaches 233 meters, whereas in the Pachelma basin, near the village of Kaverino it decreases to 198 meters. In the direction of the Ul'yanovsk projection of the basement, their thickness decreases to 76 meters near the village of Poretskoye and to 31 meters near the village of Bol'shoy Sundyr'. In Poretskoye, the lower Shchigry consists of sandstone, and near Bol'shoy Sundyr' fine-grained quartzitic sandstone is overlain by silty limestone with Atrypa ex gr. reticularis (Linn). On the Ul'yanovsk projection of the basement these sediments were not preserved.

As noted above, on the western slope of the Volga-Ural massif, in most sections of terrigenous Upper Givetian and lower Shchigry formations, chamositic formations are not developed. The only exceptions are lower Shchigry formations on the marginal, southwestern part of the massif near the villages Tokmovo, Chaadayevka, and Kikino. In this strip, in beds of argillaceous siltstone there is an interlayer of bright green chamosite in lenticular oolites, cemented by silty clay.

When comparing the sediments filling the Kazan-Sergiyevsk basin with those on the western slope of the Volga-Ural massif, it is obvious that they are quite different. The western slope is not characterized by Domanik type sediments, chamositic iron ore, bauxite, chamositic ferruginous clay and marl, or volcanics. The sections also differ in completeness; these are the elements of the structure of the sedimentary mantle of the Volga-Ural massif.

2. Structure of the Kazan-Sergiyevsk Basin at the End of the Kynovsk

The tectonic map (Fig. 1) shows that between the Tatar and Pokrov-Buzuluk projection of the basement there is a wide transverse depression through Tuymazy-Buguruslar and filled with pre-Devonian and Devonian sediments. Several basins filled with Devonian terrigenous sediments branch off from the Sergiyevsk depression. One of these is southwest of the Samara bend of the Volga and extends into the Saratov basin; a second, northeast of Melekes and west of Kazan crosses the Volga-Ural massif initially in a northern direction, and then in a north-northeast direction; a third is traced near Zainsk and Golyushurma along the eastern margin of the western block of the Tatar uplift; and a fourth in the region of Bakaly village

The limits of the Kazan-Sergiyevsk basin coincide with exposures of a complete stratigraphic section of "terrigenous Devonian" sediments. Due to insufficient data it is not always traced accurately.

The limit of complete sections passes in the west, north of Samarskaya Luka, through the following villages: Uzyukovo, northeast of Melekes, through Urmary, Mariinskiy, Posad, southwest of Yoshkar-Oly and Kotel'nich. West of this line terrigenous Devonian sediments are missing, and younger Devonian formations cover the crystalline basement. Similar relations occur also in the southern border of the Kazan-Sergiyevsk basin. South of this border, in the Pokrova region, terrigenous Devonian sediments do not appear. East of the Pokrova projection of the basement, where data on drillings are as yet unknown, they are probably missing or only partly developed. Geophysical investigations show show that south of the village of Mukhanovo-Pilyugino, the transverse Buzuluk projection can be traced, and north of it the basement is believed to be more than 3,000 meters deep [21]. This projection extends along the southern edge of the Kazan-Sergiyevsk basin.

The limits of distribution of the complete terrigenous Devonian section extend east of the villages of Vozhgaly, Kazaklar, Verkhniy Uslon, and Kamskoye Ust'ye. This line separates the Tatar uplift from the Kazan-Sergiyevsk basin. On the uplift, in the immediate vicinity of the basin, almost all the terrigenous Devonian sediments are missing, except a thin layer of Kynovsk, several meters thick, covering a part of the projection. Southeast of Kamskoye Ust'ye the limits of the depression extend to the southeast, and in the Aksubayevo region become latitudinal until they meet the meridional contours of the Birsk basin between the Tatarian and Bashkirian projections of the basement. This latitudinal outline of the Kazan-Sergiyevsk basin coincides with the clear boundary of the complete section of terrigenous Devonian formations and extends into a zone of thicker pre-Devonian volcanic rocks.

The data cited show that in the Kynovsk, the Kazan-Sergiyevsk basin was a narrow, graben-like structure, with a latitudinal branch 200 kilometers wide, and a meridional branch, probably not wider than 90 or 100 kilometers. The apparent length of the basin exceeds 800 kilometers.

The tectonic map shows the thickness of various units of the terrigenous Devonian inside the basin and on the adjacent structures. These data indicate that, at the end

of the Kynovsk, the Kazan-Sergiyevsk basin was filled with sediments 800 meters thick in the deepest part; the different stages have various thicknesses. In the Sovetsk region, the thickness of Middle Devonian, Pashiysk and Kynovsk sediments reaches 620 meters, and in the southeastern part of the basin their thickness does not exceed 270 meters. Near the village of Bavly, pre-Devonian sediments are 540 meters thick, and near Buguruslan more than 600 meters. Between the most warped northern part of the basin and its southern branch, the thickness of the terrigenous Devonian sediments is 203 meters, whereas older sediments, as previously mentioned, are entirely missing here.

In the basins flanking the Tatar uplift, the thickness of the sediments usually does not exceed 110 meters (Bakaly, 88 to 107 meters; Zainsk, 73 meters; Golyshurma, 85 to 104 meters). On most of the Ul'yanov-Chuvash projection of the basement; these sediments do not appear. The thickness of terrigenous Devonian formations also varies; near the village of Borovka, 263 meters; near the village of Mukhanovo, 210 meters; in the Sterlibashevo region, 163 to 176 meters. These data show that, at the end of the Kynovsk, in the latitudinal branch of this structure, there was, between Zol'nyy-Ovrag and Sterlibashevo, an elevated section higher than the Radayevka section, similar to the part of the basin extending northwest from the village of Radayevka in the Kamskoye Ust'ye region. The Sterlibashevo region is somewhat higher than the Zol'nyy Ovrag region. The variation in the thickness of the rocks in the latitudinal branch of the structure is measured in tens of meters. The most significant change in thickness, 415 meters, is between the meridional part of the Kazan-Sergiyevsk basin and the relatively elevated district of Kazaklar-Kamskoye-Ust'ye.

3. Volcanic Manifestations in the Kazan-Sergiyevsk Basin and their History

In the zone connecting the Kazan-Sergiyevsk basin with the relatively elevated region south of the Kazaklar, there are extrusives 45 meters thick; they form the upper part of the Pashiysk section. The lavas and tuffs are covered by sediments containing Kynovsk fossils [17]. The extrusive cover is not homogenous; it consists mainly of vitreous and partly crystallized rocks and is covered by ashy tuff. B.A. Uspenzkiy [18] noted that the tuff is composed of peculiar particles, similar to drops of lava, 0.1 to 0.6 millimeters in diameter. Clastic quartz silt is also found in minute quantities. The vitreous substance is not crystallized and its colors range from light to dark brown. In places, a perlitic structure is encountered. The refractive index of the glass is somewhat higher than that of Canada Balsam. Uncrystallized varieties alternate with crystallized ones, as seen clearly under polarized light. They are characterized by a microlitic, refined texture, and consist of aggregates of minute acicular microlites of plagioclase embedded in a vitreous mass. The texture of the rock is hyalopylitic. Some of the plagioclase microlites have a flow texture. Crystallized lavas are marked in places by accumulations of minute acicular crystals of magnetite and titanomagnetite. Textural features of the extrusives, refractive index of the glass and presence of titanomagnetite all indicate an andesitic-basaltic composition for the extrusives [18].

It is interesting to clarify the role of the Kazaklar andesite-basalt in the Devonian history of the Kazan-Sergiyevsk basin, and its relation with the older, pre-Devonian, volcanics. The filling of the Kazan-Sergiyevsk basin began already in pre-Devonian times when the Bavly formations filled the region of Zol'nyy Ovrag, Radayevka and Bavly villages. To the south and east, these depressions merge with the Sergiyev depression, its southern side having, according to geophysical data, a linear form, roughly at the latitude of Mukhanovo-Sterlibashevo. Along the borders of the ancient Sergiyevsk depression, pre-Devonian volcanics occur among crystalline rocks, and occasionally are included in the lower Bavly formation (Fig. 1, 13). In the southern part of the depression, gabbro and norite occur among crystalline rocks near Zol'nyy-Ovrag, Pecherskoye and Syzran'. In the inner part of the depression, diabase is exposed, near the village of Shkapovo. Outside the depression, veins and analogous extrusive basic rocks are located along its margin, passing along the eastern boundary of the Tatar uplift, near the villages of Chekmagushy, Or'yebashevo, Kaltasy, Varzi-Yatchi, Pychas and Grakhovo, and also along the western boundary near the villages of Kazaklar and Mariinskiy-Posad [2, 13].

The formation of the Kazan-Sergiyevsk basin continued in the Devonian, in the extrusive zone of the Riphean depression. This began with the accumulation of Takata sandstone, 20 to 30 meters thick, along the southern margin of the basin. In the Early Givetian the basin extended more to the north and northeast, and it began to receive sediments containing marine fossils. The largest accumulations of carbonates and argillaceous carbonates occurred again along the southern margin, where the Lower Givetian formations, near Mukhanovo-Sterlibashevo, are 30 to 50 meters thick; near the northern boundary of the basin, near the Chekmagushy, Bakaly and other villages, the Lower Givetian does not exceed 10 meters in thickness. Also formed was a narrow basin near Sovetsk on the arch

of the Volga-Ural crystalline massif. The eastern and western boundaries of this basin were probably determined by the location of Riphean volcanics at Mariinskiy Posad and Kazaklar.

The Kazan-Sergiyevsk basin reached its maximum development in the Late Givetian. Its northern branch (Sovetsk region) merged with the latitudinal Sergiyesk pre-Devonian, Eyfelian and Early Givetian depression. This resulted in a single synclinal form, the Kazan-Sergiyevsk basin. In the region of Sovetsk, almost 285 meters of Upper Givetian sediments accumulated. South of this basin, the center of the depression was shifted towards the northern edge of the basin in the region of Radayevka, Borovka and Sernovodsk, where the upper Givetian terrigenous and calcareous rocks exceed 130 meters in thickness. Towards the end of the Upper Givetian, volcanic activity occurred in the basin, as indicated by the tuffs penetrated in drillings in southeastern Tatariya [12].

In Pashiysk time, sandstones bearing carbonaceous detritus continued to accumulate in the Sovetsk depression, while mainly argillaceous silty sediments accumulated to the south, with interlayers of limestone containing brachiopods and lingulas, and also fragments of fish bones and plant remains. During the same period another depression formed near the village of Zol'nyy Ovrag where the arenaceous and argillaceous sediments reach a thickness of 100 meters [14], whereas near Radayevka they do not exceed 65 meters. During Pasha time, lava was extruded near the village of Kazaklar, where the thickness of the lower and upper Givetian and Pasha sediments changes abruptly. This is shown by the tuffs of the Radayevka region, which occur in the sediments of the southern branch of the basin. The Kynovsk sediments marked still further the Sovetsk and Radayevka depression from the neighboring structures. From the end of the Kynovsk till the end of the Devonian, sediments consisted almost entirely of calcareous rocks, covering the entire Tatar uplift, the Kazan-Sergiyevsk basin, on the Ul'yanovsk-Chuvash projection of the basement, and also on the western slope of the massif. The sharp boundaries of the Kazan-Sergiyevsk basin began to fade during Sergiyevsk time, and in the southwest, the eastern boundary of thick Sargayevskiy sediments was located along the meridian of the village of Baytugan [16]. In the Middle Frasnian, the outlines of the basin changed still further, and in the Upper Frasnian and Famennian the basin was no longer an integral structure. It is interesting to note that this coincides with the disappearance of volcanic and ferruginous-chamositic terrigenous formations from the section.

Data touching the relation between the diabase and tuff, and the ferruginous oolitic strata on the Timan, are of interest. Among arenaceous and argillaceous rocks of the Yarega formation of the Middle Devonian, on the middle Timan and in the basin of the Ukhta River, there appear dikes and formations of tuff and diabase. D. P. Serdyuchenko [15] showed that in the upper reaches of the Izhma River the tuff-diabase unit coincides stratigraphically with the 35 meter thick "iron-ore" bed. This unit is represented by various terrigenous and calcareous rocks, with tuff and "iron-ore" interlayers consisting of ferruginous and kaolinitic clays containing goethite oolites and ferruginous chlorite, associated with siderite. The tuff and ferruginous sediments gradually merge into one another. The light and heavy fractions of the "iron ore" contain considerable amounts of volcanic glass, and the diabase tuff contains many interlayers and lenses of oolitic-argillaceous-hydroxide iron ore, sporadic concentrations of oolitic-ferruginous chlorite, and grayish-brown kaolinitic-goethitic oolites.

The mineralogic and chemical similarity of the ferruginous-chloritic formations, in the Kazan-Sergiyevsk basin and on the Timan, their similar geologic location and spatial connection with volcanics, all warrant the assumption that there exists a paragenetic connection between the andesitic-basaltic lavas, tuff and ferruginous-chamositic formations in the Kazan-Sergiyevsk basin.

CONCLUSIONS

The history of the origin and development of the Pachelma basin shows that, when a sedimentary mantle began to cover the Russian platform, graben-like structures began to form [22]. The study of the Pachelma, Kresttsy, Kazan-Sergiyevsk and other basins shows that the development of tectonic forms of the sedimentary mantle of the platform began by the breaking up of the basin into large blocks, and the formation of graben-like basins between them. In Middle Devonian time the graben-like Riphean Pachelma basin developed into a broad synclase [22], and the Kresttsy and Kotlas Riphean basins were buried under the neighboring Moscow syncline.

The tectonic development of the Kazan-Sergiyevsk basin in Middle Devonian and Early Frasnian times was different. It continued to develop along the lines of the Riphean basin, wedging into the body of the Volga-Ural massif. The basin grew by "chipping" of the adjacent parts of the old massif, and also by the subsidence of new areas. At the beginning of the Middle Frasnian it lost its graven-like appearance, probably due to the change in tectonic form,

coincident with the cessation of volcanic activity.

Volcanic rocks of Riphean and Devonian age are distributed in the marginal areas of the Kazan-Sergiyevsk structure where it adjoins the basement. The marked permeability of lavas of the marginal parts is characteristic also of the Riphean basins of the central part of the Russian platform. Magmatic formations testify to the disruption of the continuous layers of sediments, accompanied by structural deformation. The Kazan-Sergiyevsk basin is a rudimentary graben-like form, whereas fracture of the continuous layers was accompanied by dislocations of fault-like character, often with an imperceptible shift of these blocks; these movements did not disturb the flat-lying beds within the blocks which were not complicated by deformation. The basins in the central part of the Russian platform are essentially similar.

Comparative study of isopach maps, constructed for various stages in the development of the Kazan-Sergiyevsk basin, shows that it is asymmetric. This is probably also characteristic of other graben-like structures. The total thickness of the Riphean and Devonian sediments filling the basin ranges probably, within 1,000 meters. Such sediments in other ancient basins in the central part of the Russian platform have a comparable thickness.

The composition of the magmatic formations in the rudimentary graben-like structure is comparatively uniform. The intrusives belong to the gabbro group, and the extrusives to their equivalent: diabase, dolerite, basalt, andesitic-basalt and their pyroclastic analogues.

The analysis of the interrelation between the rocks filling the Kazan-Sergiyevsk basin and adjacent structures shows that the ferruginous-chamositic ores, bauxitic and chamositic-ferruginous clays, and also the terrigenous-calcareous rocks, form an association tightly connected spatially with the marginal parts of the basin. The tuffs and andesitic basalts are also closely related to this structure. These interrelations disappeared together with the disappearance of the graben-like structure. Further studies will enable us to distinguish geologic formations characteristic of graben-like basins and adjacent uplifts.

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Geological Institute,
Academy of Sciences, U.S.S.R.
Moscow

Received January 24, 1959

STRUCTURAL FACIES ZONES OF THE LOWER CAMBRIAN AND RIPHEAN AT THE SOUTHEASTERN MARGINS OF THE SIBERIAN PLATFORM[1]

by

V. V. Khomentovskiy

Present concepts about the Sayan-Altay fold region, as given in V.A. Kuznetsov's work [11], represent an important synopsis of a period of intensive study. V.A. Kuznetsov presents a detailed and concise tectonic analysis of the region based on its main geologic and structural features.

The next stage of investigation deals with the delineation of structural facies zones, especially of the older formations. This makes possible a much more elaborate tectonic regional subdivision, including genetic classifications of the elements involved.

The first attempts in this direction were made quite some time ago, by S.V. Obruchev [15] and A.S. Khomentovskiy [22], who distinguished three different types of section in the Lower Cambrian in eastern Sayan. V.M. Yaroshevich [26] noted the presence of Lower Cambrian geosynclines and geanticlines in the Kuznetsk Alatau. The structural facies of Gornaya-Shoriya were described by Yu.G. Shcherbakov [25]. However, the division into structural facies zones had to be put off until recently because of the slow processing of the biostratigraphic data.

As a result of the field studies from 1953 to 1958, I.T. Zhuravleva, L.N. Repina, and the author succeeded in determining the following stratigraphic sequence in the Lena and upper Aldan stages (starting with the oldest): the Kameshkov; Bazaikh; Sanashtykgol'; Solontsov and Obruchev units. These units were compared to the standard stratigraphy of the Lower Cambrian of the Siberian platform [5]. The results of the biostratigraphic study permitted a reliable correlation of the sections, and on this basis, a new approach to the delineation of structural facies was worked out.

The formation boundaries shown in the northeastern part of the area in Figure 1 had been verified earlier [23]. Here, the Lower Cambrian platform type is outstanding, characterized by its thickness and the absence of strong folding. Deposits of this type occur in the extreme west of the Yenisey Range and in the southern Sayan region. The boundary of the platform-type area has been modified since the first version of the map [23] on the basis of data provided by M.A. Semikhatov [20].

To the southwest, the Lower Cambrian platform-type beds are replaced by beds of approximately the same composition, but notably thicker. As a rule, they are considerably deformed and, in places, penetrated by intrusions. The presence of flysch-like deposits is characteristic, and extrusives are practically absent. Chronologically, these beds appear at almost all stages of the Lower Cambrian.

This type of section is, as previously noted [23], characteristic of the extensive miogeosynclinal foredeep which bordered the Siberian platform on the southwest, from the northern spurs of the Yenisey Range to Lake Baykal. In the miogeosyncline, as on the platform, the Cambrian strata are bedded on the Riphean with marked unconformity. The border between the shield and the miogeosyncline passes along an acute flexure. Along this line, on the Yenisey Range [10], substantial changes take place in the Riphean section. East of this line, the bulk of deposits consists of carbonate beds, while to the west terrigenous and extrusive formations predominate. The Riphean of the eastern Sayan conforms to the latter type. Further southwest, the miogeosynclinal facies of the Lower Cambrian section is replaced by the eugeosynclinal facies, characterized by extrusive formations of great thickness, in places intensively metamorphosed and penetrated by numerous intrusions.

The east Sayan anticlinorium, composed of lower Proterozoic formations, constitutes the border between the mio- and eugeosynclines. This is, apparently, a narrow tectonic

[1] Strukturno-fatsial'nyye zony nizhnego Kenbriya i rifeya yugo-zapadnogo obramleniya Sibirskoy platformy.

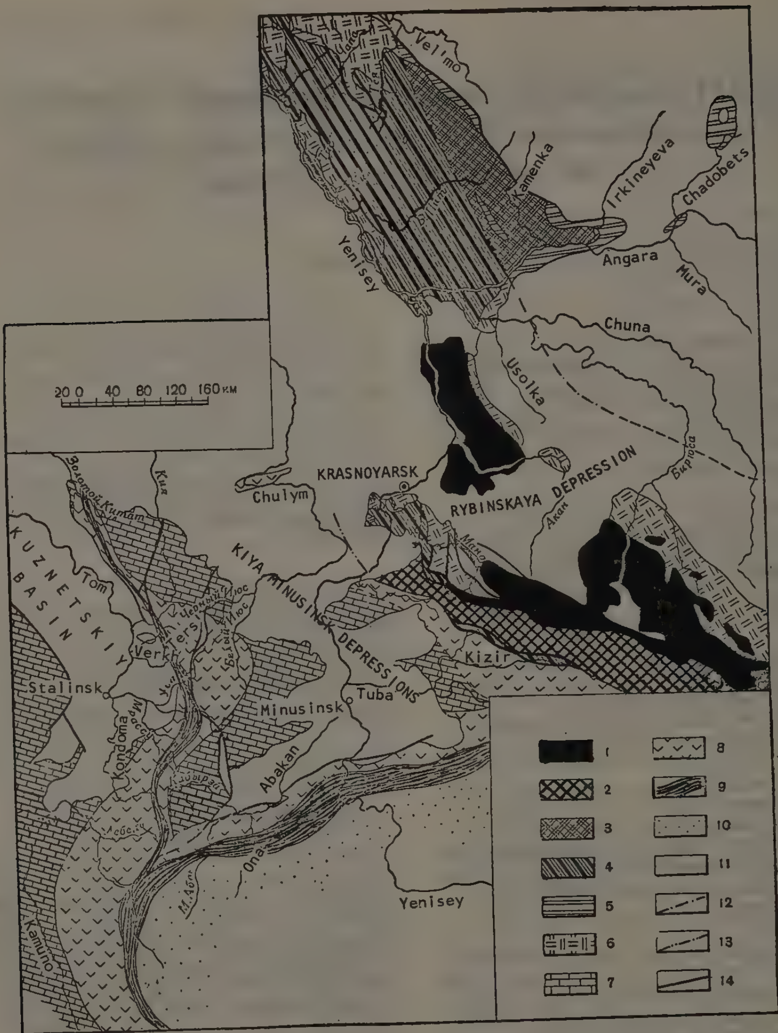

FIGURE 1. Outline of the structural facies zones of the Lower Cambrian and Riphean of the southeastern margins of the Siberian platform.

1 -- Archean (?) strata, forming the basement of the Cambrian miogeosynclines; 2 -- lower Proterozoic strata of the eastern Sayan, forming the Lower Cambrian uplift along the border of the eu- and miogeosynclines; 3 -- upper Proterozoic, mainly calcareous strata of the Yenisey Range; 4 -- upper Proterozoic (on the Yenisey Range and the Penchenga series), terrigenous, carbonate and extrusive formations; 5 -- the platform type of the Lower Cambrian; 6 -- the miogeosynclinal type of the Lower Cambrian; 7 - 9 -- the eugeosynclinal type of the Lower Cambrian and the Riphean; 7 -- mainly carbonate formations; 8 -- mainly extrusive formations; 9 -- deeply metamorphosed regions with belts of high-permeability formations; 10 -- Ordovician deposits of the western Sayan; 11 -- post-Middle Cambrian deposits; 12 -- boundary of the Siberian platform, Lower Cambrian; 13 -- boundary of the mio- and eugeosynclines in the Lower Cambrian; 14 -- principal faults (dotted sections denote inferred position).

block which rose as a cordillera above sea level during the entire Lower Cambrian. A similar point of view is expressed in the explanatory notes to sheet N-46 (Abakan) of the official geologic map [16].

In the eugeosynclinal part of the Sayan-Altay fold region, two distinct types of Lower Cambrian-Proterozoic sections are clearly discernible, one essentially carbonate and the other essentially extrusive. The area most typical of the former is the Batenevskiy Range. Here the entire Lena stage and the lower part of the Middle Cambrian are composed almost completely of carbonate deposits. Downwards, the section includes beds of greater thickness (the Yenisey series of A.N. Churakov [24], which conform chronologically to the Aldan stage and possibly, in part, to the Riphean [5]. An almost analogous section is found in the Bellykskoye Belogor'ye (data of Ye.A. Shneyder, L.N. Repina, and others). Farther east, the carbonate type is found near Kameshki village [23] and along the Kazyr River near Tayaty village, where the middle part of the limestone formation contains remnants of archaeocyathids of the Bazaikh type.[2]

The Lower Cambrian section of western Sayan, where the Aldan and Lena stages are composed almost entirely of extrusives are in marked contrast with the former [5]. The dividing line between the two sections is clearly drawn. The northeastern limit of the carbonate type is much less distinct. However, thick beds of Lower Cambrian extrusives between the Middle Cambrian and the Riphean, found along the Sisim River, as well as other indirect evidence [23], permit us to recognize the presence of a narrow strip of the extrusive section between the extensive carbonate area and the eastern Sayan anticlinorium. Along this strip rises the Arga Ridge, whose Lower Cambrian formations are also essentially extrusive.

The carbonate section can be traced from the Batenevskiy Range to northwest of the Kuznetsk Alatau and the Kiya Basin [6]. There the Aldan stage of the Lower Cambrian (Bel-Su series), as well as the Bazaikh and apparently the main part of the Sanashtykgol' units of the Lena stage are composed of limestone. According to investigations made in 1958 by the author, together with L.N. Repina and I.T. Zhuravleva, this lithology can be traced from the Batenevskiy Range southwest to the Mrassu River basin. Here, evidence of archaeocyathids and trilobite fossils enables us to relate the limestone strata to the Bazaikh, Sanashtykgol' and Solontsov units of the Lena stage. Downwards, this section is gradually replaced by a thick carbonate layer (west-Siberian and other series of K.V. Radugin [18]) which conforms chronologically to the Aldan stage and, apparently, to part of the Riphean.

The western boundary of the typical carbonate section passes within the limits of Kuznetsk Alatau and Gornaya Shoriya. The carbonate beds of the Lower Cambrian near Malaya Syya Village (on the Belyy Iyus River), as pointed out by V.M. Yaroshevich [26], are replaced by thick extrusive formations. V.M. Yaroshevich's contention was sharply criticized at the Leningrad conference (January, 1956) on standardization of the stratigraphy of Siberia. Since only one single, Middle Cambrian unit in the whole section was fossiliferous, many investigators tended to view the section in a reverse order. Our finding of Lower Cambrian trilobites [5] in strata underlying Middle Cambrian formations, vincidates V.M. Yaroshevich's point of view.

According to our findings, the typically extrusive section can be traced westward without interruption from Malaya Syya village to the eastern edge of the Kuzbass. In the basin of the Usa and Chernyy Iyus rivers, schistose extrusive layers, interbedded with limestone, form the Aldan and Lena stages of the Lower Cambrian [5]. To the north and south of this section the extrusive strata of the upper Riphean and Lower Cambrian can be traced, as a considerably wide band, along the entire length of the Kuznetsk Alatau. Along the Zolotoy Kitat River, in the northern Kuznetsk Alatau, the carbonate section of the lower Lena stage of the Kiya River is replaced by a thick formation of extrusives, interbedded with limestones containing archaeocyathids of the Bazaikh unit (data of G.F. Gorelov). On both the Zolotoy Kitat and the Kiya Rivers, the deposits of the Lena stage are underlain by the so-called Bel-Su series, but while along the Kiya River this formation consists entirely of limestone, along the Zolotoy Kitat River it consists of 30 to 70% of extrusive rocks.

A sharp facies change occurs west of the mouth of the Kabyrza River in the Gornaya Shoriya and in southern Kuznetsk Alatau. The carbonate beds of the Lena formation, which are very extensively developed in the east, are replaced here by the so-called Mrassu series [18] of an extrusive and terrigenous sedimentary composition [25]. The thick carbonate formation which underlies the limestones of the Lena stage is also changed considerably. West of the Mrassu valley near the town of Spassk, as well as in the Antrop River basin, it contains large amounts (up

[2] Here and elsewhere in this article, where no mention is made of who examined the fossil, it is to be assumed that the trilobites were investigated by Repina and the archaeocyathids by Zhuravleva.

to 50%) of schist and extrusives. In the axial region of the Kuznetsk Alatau the section of the Lower Cambrian and Riphean is also of the extrusive type, forming a strip which extends southward to the Yumeno-Lebedskoye synclinorium of the Gornyy Altay. There it consists of thick extrusive-schist series which conform chronologically (latest fossil data by O.K. Poletayev, V.M. Sennikov, A.F. Belousov and others) to the Lena and, apparently, to a considerable part of the Aldan stage.

Farther west, the extrusive schist series of the Lower Cambrian and Riphean become, once again, largely carbonate. This change is most pronounced in the Gornyy Altay, at the juncture of the Uymeno-Lebedskoye synclinorium and the Katun anticlinorium. There the section is predominantly carbonate, comprising the Riphean (data by M.K. Vinkman), the Kameshkov Bazaikh units of the Lena stage, (archaeocyathids of the Sarasy River, and the Sanashtykgol' fauna of the Postpaul region) and Obruchev units (fossils along the Katun, near Ust'-Sema).

The above-described section of the Katun anticlinorium extends into the Salair. Farther north, the carbonate and extrusive facies are in direct contact under the Kuznetsk Basin, although thick carbonate beds of Lower Cambrian age, northwest of the Gornaya Shoriya (Mundybash region), indicate that a large portion of the Kuznetsk Basin is underlain by carbonate rocks.

Summarizing the above, we see that the inner part of the Sayan-Altay fold region consists of carbonate strata belonging to the Upper Riphean and the lower part of the Lower Cambrian, forming a giant triangle that resembles in shape the Minusinsk depression. This triangle is bordered on all sides by moderately broad bands of contemporaneously formed extrusive formations.

The Kuznetsk Alatau example shows a narrow strip of extrusives across an extensive area of uniform carbonate rocks.

The facies zones described coincide with structural zones. Firstly, the extrusive facies is of greater thickness than the carbonate facies, and forms, therefore, troughs and depressions. Secondly, the extrusive-filled depressions are marked by compressed linear folding, oriented according to the strike of the depression.

The carbonate facies regions are usually characterized by wide folds (box folds and brachyfolds), commonly having quite fantastic shapes, and various changing orientations. The structure of the facies zones is most clearly grasped when observing the peculiar bends that are found in the central parts of the depressions. These bands are closely associated with deformed, strongly metamorphosed beds, ranging in age up to the Recent; they are the subject of lively discussion. Some geologists (V.A. Kuznetsov [11]; A.L. Dodin [14]) regard them as Proterozoic or even Archean formations, while others (D.V. Nikitin [14]; T.M. Dembo) see in them the metamorphic derivatives of Paleozoic deposits.

The result of our investigations confirms the second point of view. In the basin of the Verkhnyaya Ters', Kibras and Usa Rivers, from the Belaya Usa River westward, there is a gradual increase in the degree of metamorphism in the extrusives of the Aldan stage, first to amphibolites, then to amphibolite gneiss. Detailed investigations have refuted A.L. Dodin's contention that an angular unconformity exists between Proterozoic and Cambrian formations in this region as indicated by the presence of a basal conglomerate in the lower part of the Cambrian. The metamorphic rocks in the region considered extend northwards in a narrow (2 to 5 kilometer-wide) strip up to the extreme tip of the Kuznetsk Alatau (the Zolotoy Kitat River basin). Southward, the metamorphic outcrops increase in exposure, forming the so-called Tomsk Massif.

It should be noted that the metamorphic and crystalline series in the Tomsk Massif appear only as narrow strips, wedged between large intrusions. These intrusions were formerly considered Precambrian, but absolute-age determinations revealed them to be probably Caledonian. According to our and, independently, Yu.G. Shcherbakov's map of the area around the Khomutovka River mouth (Mrassu River basin), one of these metamorphic strips consists of metamorphosed Lower Cambrian deposits, which were found to contain archaeocyathids in the vicinity of Sredniy Chilay village, somewhat to the south. South of the Tomsk Massif the metamorphic outcrops decrease again in exposure, and may be followed as a narrow strip through the Gornaya Shoria to the Gornyy Altay.

The areas south of the Tomsk Massif were thoroughly mapped by Yu.G. Shcherbakov and V.I. Fominskiy. Because of the increasing southward-progressing burial of the structure, the intensely metamorphosed series do not belong to the Riphean and Aldan stages here, but to the Lena and possible the Middle Cambrian. A narrow arch of metamorphic schist connects the Tomsk Massif with the wide band of strongly metamorphosed, Paleozoic formations (N.P. Nekhoroshev [13]), which borders the Uymeno-Lebedsk synclinorium on the east.

Another branch, also connecting with the

wide metamorphic belt of the Gornyy Altay, passes along the axial part of the western Sayan (Dzhebash upwarp). The essential similarity of the metamorphic rocks of both structures has been convincingly shown by I.N. Kazakov [8]. The Dzhebash upwarp is bordered by great faults along most of its length, and therefore, no direct data is available on its green schist. Some investigators (I.N. Kazakov, A.G. Sivov, G.G. Semenov) relate them to the upper Precambrian, while others (S.A. Salun) tend to regard them as metamorphosed Cambrian formations. Our 1958 investigations in the uplands of the Abakan River convinced us that the Dzhebash series is intimately related to Lower Cambrian formations (Monok series). Evidence to this effect is seen in the sharply increased metamorphism of the Lower Cambrian formations when approaching the Dzhebash upwarp; a great similarity in composition (even primary) between Lower Cambrian rocks and many rocks of the Dzhebash series; and a similarity between the inner structure of the Dzhebash upwarp and the northern branch of the western Sayan. One is led to believe that, though the Dzhebash series appear to be mainly Riphean, it includes some younger formations. V.V. Bogatskiy, who came to similar conclusions, even believes that some units of the Dzhebash series belong to the Ordovician Shignetsk stage.

Our map shows clearly how the zones of intensive metamorphism coincide with the central areas of the geosynclinal troughs. The metamorphic belts even widen and contract in accordance with the breadth of the troughs. On the whole, they form a sort of frame outlining the structural facies zones of the eugeosyncline.

The intensely metamorphosed belts are bordered and warped by tectonic deformations almost along their entire length. Moreover, they apparently represent deep fractures. Evidence to this is seen in their vast length, each branch exceeding 700 kilometers, and in their association with magmatic hearths, which might explain the extent of intensive metamorphism over such large distances. Numerous ultrabasic intrusions are also connected with these belts.

Another piece of evidence for placing these tectonic phenomena in the deep fracture category is their antiquity. Being the chief genetic factor of the geosynclinal troughs, these fractures must have been in existence at least since the late Proterozoic. In the Kuznetsk Alatau, these fractures existed as a zone of continuously progressing metamorphism until late in the Cambrian. After that, during the entire Paleozoic, it manifested itself by intense crushing along the rigid block which the area had now become. In the western Sayan and the Gornyy Altay, the first stage of fracturing persisted up to and throughout the Ordovician.

Thus, we see that the deep fractures under consideration are peculiar in the fact that they do not form a plane or a system of planes, along which vast displacements occurred, but occur as zones of weakness, highly penetrable to metamorphosing solutions. The formation and development of such fractures determined the tectonic structure and consequent development of the whole region. During the Riphean, geosynclines were formed, which were later filled in with volcanic material. During the Lena stage, these trenches reached their greatest breadth. By the end of the Lower Cambrian, the distribution of facies had become highly diversified. The massifs had been shattered by a fault system associated with the deep fractures, and became principal areas of sedimentation of developing carbonate series. Parallel to this, extensive lava sheets flowed over increasing areas, up to the Middle Cambrian, when they almost completely covered the area now being investigated.

The folds of the carbonate strata of the massifs have a strike ranging between near-meridional, NE, and NW. This feature is, in our opinion, to be attributed to the subsequent influence of the aforementioned deep fractures, rather than to the contours of the platform (N.S. Zaytsev, [7]).

Faults occur at the folds in the high-penetrability zones. These faults are analogous with faults that accompany re-entrant angles, (N.S. Shatskiy, 1946). This is the case in the Balyksin, Saralin, Tydyn-Kundat and other fault zones which border the high penetrability belt of the Kuznetsk Alatau. Therefore it is not surprising that the folds caused by these fractures strike obliquely towards the structural facies zones.

In conclusion, we see that the deep fractures of the Kuznetsk Alatau ceased to function as unstable high-penetrability zones by the Upper Cambrian. At approximately the same time, the geosynclinal processes died down in these regions, while in the western Sayan, which has been recognized as a Caledonian structure, the deep-fracture system continued to function well into the Ordovician as a high-penetrability zone. All this is evidence that the high-penetrability zones constitute independent structural elements, playing a highly important role in geosynclinal formation.

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Geological Institute,
Academy of Sciences of the U.S.S.R.
Moscow

Received March 6, 1959

THE MOST IMPORTANT FEATURES OF QUATERNARY (ANTHROPOGENE) DEPOSITS IN THE NORTHEASTERN PERIPHERAL PART OF THE CHUYSK TROUGH[1]

by

V. I. Yeliseyev

1. Introductory Remarks

The Quaternary geology of the northeastern peripheral part of the Chuysk trough[2] is in its initial stage. Quaternary deposits have been studied here, quite inadequately, in conjunction with hydrogeologic study and with prospecting for diamonds. Their study is also hampered by a number of factors, such as the absence of a mammalian fauna. Over large areas, Quaternary alluvium is now covered by aeolian sand and is inaccessible to direct study. Its study will call for deep drilling because the overburden in this area is fairly thick.

The paleoclimatic method, widely used in differentiation and correlation of Quaternary deposits in glacial regions, loses some of its effectiveness, here. The geomorphic method, although important, cannot be fully applied everywhere because the terraces are poorly expressed in certain provinces having a general tendency for subsidence (such as the Chuysk trough), especially in zones of recent subsidence. In addition, these terraces have been strongly deformed by aeolian processes. Therefore, the dating of Quaternary stages, as given below, is approximate and tentative. Only the sequence of Quaternary deposits is determined with certainty, by the complex method.

2. The Lower Boundary of the Quaternary (Anthropogene) System

Most students of the Quaternary (V.I. Gromov, A.I. Moskvitin, N.I. Nikolayev, K.V. Nikiforova, etc.) believe that its lower boundary should be moved downward, to include the Upper Pliocene. Field data from the northeastern peripheral part of the Chuysk trough, as well as published material on the neighboring regions of Kazakhstan and Central Asia, confirm the necessity of pushing back the lower boundary of the Quaternary, to the base of the upper Pliocene continental deposits.

Upper Pliocene deposits of this area (the Kenshagyr formation) are marked by their generally straw to gray color, similar to that of the Quaternary and different from older deposits which display red hues. This is undoubtedly due to the cooler upper Pliocene climate.

Upper Pliocene cooling is also noted for other areas of Kazakhstan and Central Asia. B.A. Fedorovich [12] notes in his analysis of the paleogeography of Central Asian plains that redbeds persist in the Sub-Akchagyl' petroliferous formation while they are missing in the Akhagyl' formation.

Great climatic changes in the beginning of the upper Pliocene are also known from Tyan'-Shan'. According to S.S. Shul'ts [15], redbeds above the straw-colored formation are missing in the area of Serafimovka village, in the Chuysk trough; the straw-colored formation he believes is late Tertiary or early Quaternary in age. Actually it is upper Pliocene because it is here that V.S. Bazhanov has collected remains of Equus stenonis [18].

In northwestern Fergana, the straw-colored formation with Elephas ex. gr. meridionalis differs from the underlying brown formation (represented by red-brown clay and sandstone) by the absence of red intercalations. A tooth of Elephas ex. gr. meridionalis dates the enclosing beds as upper Pliocene, according to Ye.I. Belyayeva [1].

Lithogenetic and tectonic analyses, too, suggest the desirability of lowering the Quaternary boundary. Straw-colored formations in the Serafimovka area of the Chuysk trough and in northwestern Fergana are made up of thick alluvial fans, considerably coarser than sediments of the underlying formations; this, in

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[1] Glavneyshiye osobennosti chetvertichnykh (antropogenovykh) otlozheniy severo-vostochnoy pribortovoy chasti chuyskoy vpadiny.

[2] This region, immediately adjacent to Chu River in the north and south, is sometimes known as the "Chuysk Steppes."

turn, suggests a more intensive mountain erosion which was obviously related to the onset of an intensive uplift in the Tyan'-Shan' area.

The upper Pliocene Kenshagyr deposits in this area are marked by quite different quantitative ratios of heavy minerals, compared with older sediments. Present here along with stable minerals are such unstable ones as pyroxenes and amphiboles, almost wholly missing in older sediments and typical of the younger ones.

Data on hand on mammalian fossils in neighboring regions also suggest a lowering of the Anthropogene boundary. Representatives of older animal groups, with a few exceptions, disappear in the upper Pliocene of eastern Kazakhstan, and are replaced by elephants and the one-toed Equus stenonis. A fossil assemblage, whose principal representatives are Equus stenonis and Anancus arvernensis, has been designated as the Ili, by V.S. Bazhanov and N.N. Kostenko,[3] and is correlated by them with the Khaprov assemblage of V.I. Gromov [2]. As they correctly state, the Ili assemblage marks the onset of the Anthropogene.

Deposits with this fossil fauna are widely developed throughout the world (Villafranchian of Italy, Upper Siwalik of India, Nihevan of China, etc.). On recommendation of the 1948 International Geological Congress in London, they are believed to be Quaternary by almost all foreign students.

Thus, on the basis of determinations by a joint method (paleoclimatic, lithogenetic, tectonic, and biostratigraphic), the lower boundary of the Quaternary should be moved back, to the base of the upper Pliocene Kenshagyr formation (or Eo-Pleistocene, according to V.I. Gromov's classification, here adopted), in the Chuysk steppes; to the base of the Ili formation, in the adjacent regions of Kazakhstan; and to the base of the straw-colored formation in Tyan'-Shan'.

3. Substantiation of Quaternary Stratigraphy in the northeastern Peripheral Part of the Chuysk Trough

The Quaternary of the Chuysk steppes (including Eo-Pleistocene Kenshagyr formation) is represented exclusively by continental sediments divisible into the following groups or series: 1) channel deposits of two genetic types, alluvium and alluvial fans; 2) chiefly lacustrine deposits; 3) aeolian deposits, represented here a single genetic type of aeolian sands; 4) salt- and mut-flat deposits; 5) eluvial-glacial deposits of mixed genetic types.

Alluvial sediments are widely developed along the left bank of the Chu where they form the large Muyunkum sandy massif. Alluvial fans extend in a wide belt (up to 16 kilometers) at the foot of the Chu-Ili Mountains.

A study of alluvial fans in ravine sections from the mountains to the Chuysk trough shows three definite terraces in the upper courses of some of them (e.g., the Sunkar ravine, near the Khan-Tau Mountains), with the lower terrace in contact with deposits bearing a modern fauna (Fig. 2). Sediments of the uppermost terrace are represented by typical conglomerates, similar in some details to the "Upper Gobi conglomerate" from Pogranichnaya Dzhungariya, described by V.A. Obruchev [10] and assigned by him to the early Quaternary.

Upper Gobi breccia-conglomerate of the Chu-Ili Mountains, underlain by the Eo-Pleistocene Kenshagyr formation, is traceable in the Dzhungari Ala-Tau area. On the south slope of that range, in the valley of Koybyn, it rests with a sharp unconformity on Eo-Pleistocene deposits with remains of Anancus arvernensis, according to N.N. Kostenko [6].

Similar deposits (conglomerate formation of S.S. Shul'ts; the Sokha sequence of N.P. Vasil'kovskiy, etc.) are widely distributed in Central Asia and Kazakhstan; they are unanimously believed to be lower Pleistocene.

In this paper, the Upper Gobi breccia-conglomerate is designated the Buruntau formation. It is possible to assign it to lower Pleistocene [2]: the fossil assemblage from deposits formerly assigned to the lower Quaternary by N.N. Kostenko [6] (at the base of which the Upper Gobi conglomerate commonly rests on the Eo-Pleistocene) gives substance to its correlation with the Tiraspol' fossil assemblage of V.I. Gromov. According to V.S. Bazhanov and N.N. Kostenko (see footnote 3) these fossils, collected in various localities of eastern Kazakhstan, are represented by the following forms: Equus caballus cf. mosbachensis, Hesperoloxodon antiquus, Elasmotherium Sibiricum, Rhinoceros sp., Rhinoceros mercki, Equus sp., Equus (Asinus) hindruntinus, Camelus sp., Alces latifrons, Cervus elaphus, Bison pricus sub sp., Canidae, and Mustelidae.[4]

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[3] In their paper read before the All-Union Interdepartmental Conference on the Quaternary; May, 1957.

[4] Here and elsewhere in this paper, fossil lists of B.S. Bazhanov and N.N. Kostenko are given with slight modifications by the author.

At a distance of 8 to 10 kilometers from the Chu-Ili Mountains, going toward the Chursk trough, alluvial fans no longer form terraces visible in ravines. Instead of deposits of different ages in contact with each other, younger sediments are observed here resting on the older. Such regular patterns in the structure of alluvial fans undoubtedly suggest different directions of contemporaneous tectonic movements: uplifts in the immediate vicinity of the mountains, becoming subsidences 8 to 10 kilometers away from there.

In the valley of Chu, south of Kosobazhon point, the floodplain with a modern fauna[5] is accompanied by three higher than flood-level terraces: the first, the second, and the third (Fig. 3). Traced southeast (toward the Kirghiz Range), the highest terrace consists, in the Beskatyn-Kuduk well area (about 27 kilometers north of Lugovaya station), of silt and sandy loam, changing southward to lower Pleistocene breccia-conglomerate. This makes it very probable that the third higher than flood-level terrace is lower Pliocene. Inasmuch as the deposits of this terrace make up most of the Muyunkum sandy massif area, we designate it the Muyunkum formation.

The first higher than flood-level terrace has a distinctive, bald surface, abounding in mud flats. Because of that, it is readily recognizable elsewhere. In areas to the west, it carries, according to A.L. Yanshin [16], Upper Paleolithic faunal remains; it is, therefore, upper Pleistocene. Deposits of this terrace are best known from the Saroy depression; for that reason we designate them as the Saroy formation. Inasmuch as these deposits, east of Kokterek village, change to alluvial fans, the latter also are upper Pleistocene.

Sediments of the second higher than flood-level terrace in the Chur valley, and of the second terrace in the Sunkar ravine, are most probably middle Pleistocene. We have named the former, the Shoshkaul'gen formation, after Soshkaul'gen point.

Thus Quaternary deposits (Anthropogene) of the northeastern peripheral part of the Chuysk trough are subdivided into Eopoleistocene, Pleistocene (upper, middle, and lower), and Holocene (Table 1).

These genetic types of deposits, with the exception of aeolian, saline-mud flat, and eluvial-deluvial, are considered below in their stratigraphic sequence.

4. Eopleistocene (Kenshagyr formation)[6] Q_1^1

Deposits of the Kenshagyr formation are very widely developed in this area, but they hardly ever are exposed or project from under younger Pleistocene rocks. They rest with a noticeable erosion on Andassay

[5]We have designated deposits with this fauna as the Chuysk formation.

[6]This formation has been named by K.V. Nikiforova, after the Kenshagyr ravine where it was first described.

Table 1

Differentiation of Upper Quaternary (Anthropogene) System in the Northeastern Peripheral Part of the Chuysk depression

System (period)	Division (epoch)	Stage (age)	Genetic types of deposits
Quaternary (Anthropogene) Q	Holocene Q_3		Alluvial (Chuysk formation), alluvial fans, aeolian, saline-mud flats, eluvial-deluvial
	Pleistocene Q_2	Upper Q_2^3	Alluvial (Saroy formation), alluvial fans
		Middle Q_2^2	Alluvial (Shoshkaul'gen formation), alluvial fans
		Lower Q_2^1	Alluvial (Muyunkum formation), alluvial fans (Buruntau formation)
	Eopleistocene Q_1		Chiefly lacustrine (Kenshagyr formation)

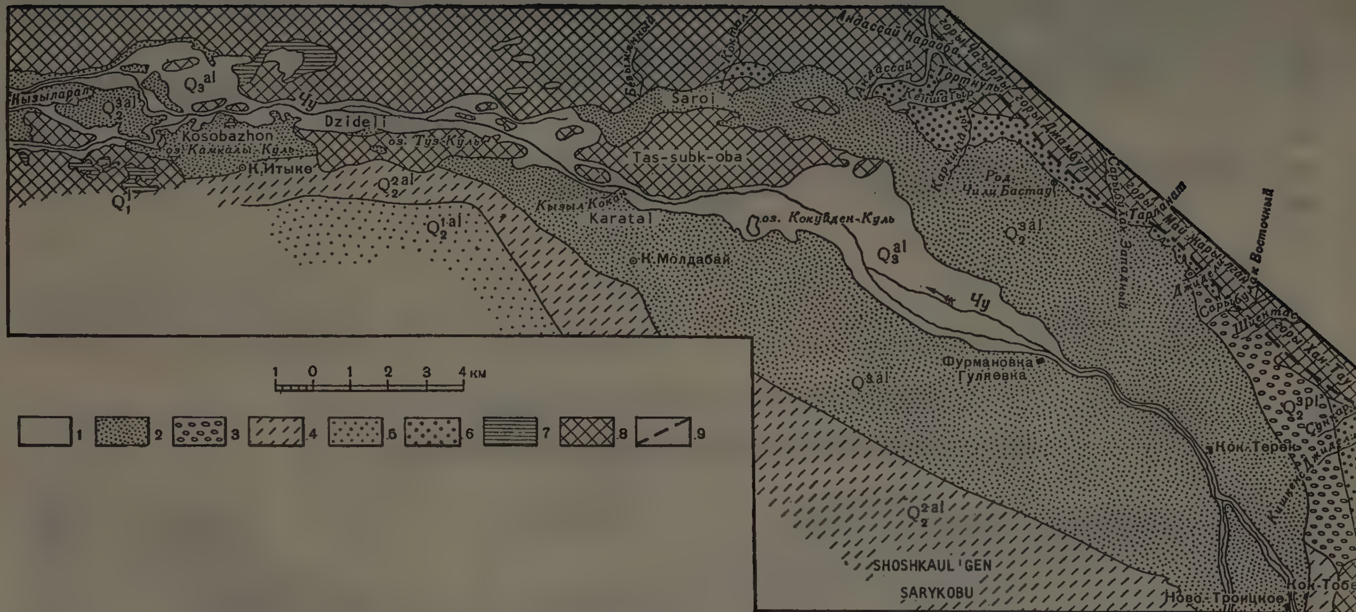

FIGURE 1. Generalized map of the distribution of Anthropogene deposits in the northeastern peripheral part of the Chuysk depression.

1 -- Holocene deposits of river valleys, ravines, and some closed depressions; 2 -- upper Pleistocene alluvial deposits; 3 -- upper Pleistocene alluvial fans; 4 -- middle Pleistocene alluvial deposits; 5 -- lower Pleistocene alluvial deposits; 6 -- lower Pleistocene alluvial fans; 7 -- Eopleistocene, chiefly lacustrine deposits; 8 -- Pre-Anthropogene rocks (eluvial on the surface); 9 -- fault traces.

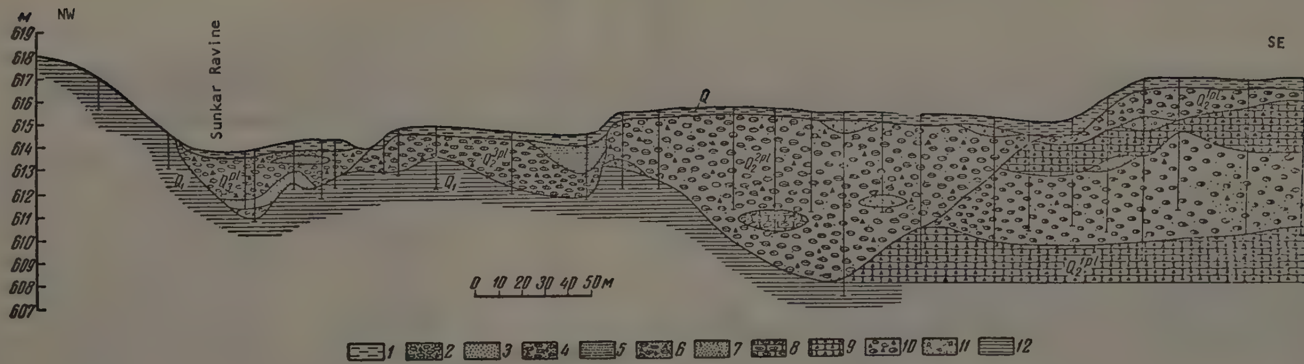

FIGURE 2. Diagram of the relation between various Quaternary deposits in the Sunkar ravine, 2 kilometers below its emergence from the Khan-Tau Mountains.

1 -- Undifferentiated Quaternary loams; 2 -- Epipleistocene rubble and gravel; 3 -- Epipleistocene sand; 4 -- upper Pleistocene rubbly gravel; 5 -- upper Pleistocene sand; 6 -- middle Pleistocene rubbly gravel; 7 -- middle Pleistocene pebbly sand; 8 -- middle Pleistocene conglomerate; 9 -- lower Pleistocene breccia-conglomerate; 10 -- lower Pleistocene rubbly gravel; 11 -- lower Pleistocene pebbly sand; 12 -- Eopleistocene clay.

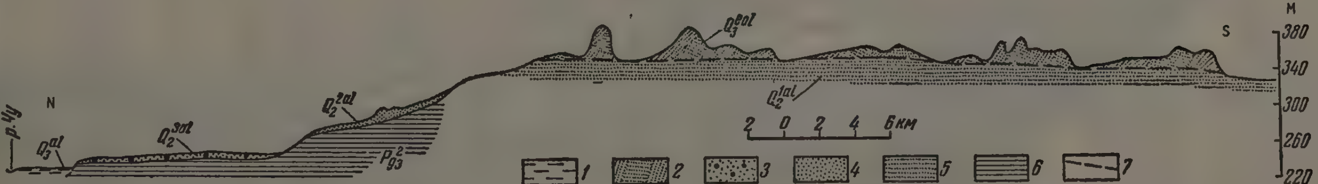

FIGURE 3. Diagram of the relation between various Quaternary deposits south of Kosobazhon point.

1 -- Holocene alluvial loam; 2 -- Holocene aeolian sand; 3 -- upper Pleistocene pebbly alluvial sand; 4 -- middle Pleistocene alluvial sand; 5 -- lower Pleistocene alluvial sand; 6 -- middle Oligocene formations; 7 -- approximate boundary between stationary and windblown sand.

Miocene clay and then on middle Oligocene, Upper Cretaceous, and Paleozoic rocks.

Predominant in them are strongly calcareous clays, straw-colored to brown; locally they are rich in poorly-rounded pebbles, with intercalations and lenses of sandstone, limestone, pebbly sand, and gravel.

Common at the base of the clays, in the Kazakh highlands, is a bed of sandstone or conglomerate, thin, and locally changing laterally to limestone with a brecciated texture. This bed is missing in some sections where the clays are in direct contact either with the Andassay clays or with older rocks.

In the Saroy depression, gray to brown gypsiferous clay, locally replaced by sand, sandstone, and conglomerate, carry ostracods Cyprideis littoralis (determined by M.I. Mandel'stam from the N.G. Shubina collection) and Ilyocypris gibba (Ramdohr) (identified by G.F. Shneyder, from my collection). The thickness of this formation ranges from 1 to 17.5 meters.

Farther east, at Tortkul' mesa, light brown clay with gravel at its base (total thickness 7.3 meters) carry ostracods Ilyocypris brady Sars and Lineocypris sp. which, according to G.F. Shneyder, suggest a Pliocene (?) age.

Southwest of the Saroy depression, the Kenshagyr formation is exposed at Karatal point (in the slope of the Kyzylkokan dry ravine) where it is represented by interbedded yellow to gray, laminated clay and gray pebbly sand, total thickness 1.6 meters. The following ostracods occur in this clay (identified by M.I. Mandel'shtam): Limnocythere ex gr. grinfeldi Liepin, L. cf. detruncata Suzin, Candona compressa (Koch), C. lactea Baird, Cyprinotus ex gr. Solini (Brady), Ilyocypris brady G. Sars, Limnocythere aff. alveolata Suzin.

In the western part of the area, southeast of Lake Bol'shoy Kamkaly-Kul', the Kenshagyr formation is made up of gray to green clay, silt, and sand, total thickness of 2.3 meters. They carry Cyprideis littoralis Br., Ilyocypris brady Sars (identified by G.F. Shneyder), also Cyprideis torosa littoralis, Limnocythere aff. alvelata Suzin. (identified by M.I. Mandel'shtam. Finally, M.I. Mandel'shtam identified the following from similar clay north of Lake Kyzylaral: Candona lactea Baird, C. cf. angulata G.W. Müller, C. neglecta G.O. Sars, C. albicans (Brady), Limnocythere aff. tenuireticulata Suzin, and Chara sp.

The Kenshagyr formation becomes thicker toward the Chuysk trough. It is 40 meters thick in the immediate vicinity of the Dzhambul Mountains (Chili-Bastau spring). Ten kilometers southwest of the Mayzharylgan Mountains (mouth of Shiyentas ravine), its partial thickness is 168 meters.[7]

Such thicknesses are a positive indication of intensive subsidence in the extreme peripheral part of the Chuysk trough, in the Eopleistocene. These are brown calcareous clays with a considerable amount of poorly rounded pebbles and rubble of indigenous rocks, similar clays, without the pebbles but with intercalations of gravel beds and limestone, and gray sand. The rubbly clays dip to the southwest at 15°.

To the south, the total thickness of this formation in the depression is not known. Outside the area, it is fully exposed near the Beskatyn-Kuduk well where it is represented by homogeneous yellow-gray to gray calcareous to gypsiferous clay with occasional intercalations of silt, total thickness, 470 meters.

According to M.I. Mandel'shtam, the above-named ostracods are upper Pleistocene (from N.G. Shubina's data). In 1959, at the author's request, M.I. Mandel'shtam kindly reexamined the ostracod collection from the Chuysk steppe Kenshagyr formation. His conclusions were the same as before. According to him, the following are typical upper Pleistocene forms: Candona angulata Müller,[8] Limnocythere aff. tenuireticulata Suz., L. aff. alveolata Suz., L. cf. detruncata Suz. A confirmation of that is found in Kh.M. Khuliyeva's statement [7] that Candona angulata Müller is an index form for the uppermost Akchagyl beds. Another typical form for upper Akchagyl beds is Limnocythere tenuireticulata Suz. [9]. In Turkmenistan, it occurs in the Lower Apsheronian but not above that [7].

Correlative with the Kenshagyr formation in the area of the northern terminal of the Karatau are deposits which carry, according to Ts.S. Grinberg, V.N. Kravchuk, and others, an Akchagyl' fauna: Limnocythere pradiginalis Mand. and L. ex gr. pliocenica Suz. [5].

Most Kenshagyr sediments probably originated in a large brackish water basin, i.e., they are chiefly lacustrine. Near the Chu-Ili Mountains, they are lacustrine-alluvial fan; in the Kazakh foothills, lacustrine-alluvial.

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[7]Thickness of Anthropogene deposits for the Chuysk trough are given as reported from boreholes of the Kazakh hydrogeologic expedition. The well cuttings and cores were examined by the author.

[8]Redetermined as Candoniella iliensis Mand., by M.I. Mandel'shtam.

Spore-pollen assemblages from Kenshagyr rocks consist almost exclusively of pollen of non-woody plant associations, with Salsola, Salicornia, Suaeda, and Artemisia in definite ascendance. Thus, wormwood-saltwort steppe was the dominant condition in the Chuysk trough, in the Eopleistocene, even as it is now. It may be inferred that the Eopleistocene climate, like the present one, tended to be dry.

5. Lower Pleistocene Q_2^1

a) Alluvial Deposits

(Muyunkum formation) Q_2^{1al}

B.A. Fedorovich states ([13], p. 211), "The Syr-Dar'ya, like its former tributary Chu, originated much later than the Amu-Dar'ya." He believes that the climate of Tyan'-Shan' where the headwaters of these rivers are located, was very dry during the entire Tertiary, with numerous closed lakes formed in isolated depressions. At the end of the Pliocene, Tyan'-Shan' became glaciated highlands and "the climate became so humid that the lacustrine depressions overflowed, with their waters running from one lake to another."

"The overflow from depressions Kara-Khodzhur, Atbash, Konurlen, Kochkor, Kyzyl-Suy, and Kemin, together with a lower Quaternary flow from the largest of Tyan'-Shan' troughs — the Issyk-Kul' — led to the origin of Chu River, then a tributary of the Syr-Dar'ya ([13], p. 211)."

Thus my own considerations as well as B.A. Fedorovich's data give reasons to believe that the Chu originated in the lower Pleistocene, in connection with the more intensive Tyan'-Shan' uplifts which brought about higher humidity in that region and a flow of Eopleistocene waters from the Chuysk trough. However, at that time, the Chu flowed considerably south of its present course, apparently filling up the deepest part of the Chuysk trough, north of the Kirghiz and the Kara-Tau ranges. Subsequently, in connection with a continuing uplift of these ranges and the adjacent part of the Chuysk trough, the river shifted to the north, leaving behind aggraded sandy plains. Characteristically, the Amu-Dar'ya and Syr-Dar'ya were shifting in the same direction, i.e., from south to north. A result of such a one-sided movement was the development of the vast sandy plains (windblown at the surface) of Muyunkum, Kara-Kum, and Kyzyl-Kum.

An idea of the composition of the deposits forming the upper part of a lower Pleistocene terrace north of Dzhidel' point is obtained from the U.M. Akhmedsafin data (1947) from the Aychek (30 kilometers southwest of the Ityk well) and the Musabek (20 km south of the Aychek) wells. These are fine-grained quartz-feldspar to calcareous sand, loam, and sandy loam, whose partial thickness at the Aychek well is 38 meters.

In the Beskatyn-Kuduk well area, the Muyunkum formation is represented by yellow-gray, locally horizontally stratified sandy loam and loam, with a total thickness of about 145 meters. The alluvium rests on clay of the Kenshagyr formation.

On the basis of the lithology of lower Pleistocene deposits (with the data not covering by far the vast area of the Muyunkum sandy massif), it may be assumed with some certainty that the contemporaneous river flow was not too strong and its gradient very low (the river was flowing over the perfectly flat plain of the former bottom of a Pleistocene lake). This stream meandered among its own deposits, often reversing its course, but shifting generally to the north. The great thickness of the alluvium was brought about by the subsidence of the Chuysk trough.

b) Alluvial Fan Deposits

(the Buruntau formation) Q_2^{1pl}

Alluvial fan Pleistocene deposits are developed in sheets at the foot of the Chu-Ili Mountains. They are either exposed or else buried under younger Pleistocene deposits, at the base of alluvial fans in some ravines emerging from the mountains. They are underlain by Miocene and Eopleistocene rocks; locally they rest on the Paleozoic. They are everywhere separated from underlying rocks by an erosional, at times angular, unconformity.

They are calcareous breccia conglomerates of poorly-rounded pebbles and rubble, from several millimeters to 3 centimeters; with fragments up to 20 centimeters. There are rough chunks, up to 1.0 meter, and occasional pockets and lenses of gray, calcareous, cross-bedded sandstone. All fractions in the breccia-conglomerate, including the large chunks, are of indigenous rocks. These conglomerates are either unstratified, or very poorly so. They range in thickness from one to five meters, near the Chagyrly-Dzhambul Mountains. Away from there, as can be seen in the Kenshygar ravine area, they change to rubble-gravel deposits, better sorted, somewhat better rounded, and not so well cemented (in places altogether unconsolidated). Appearing in them are lentils and lenses of sand, loam, and marl, with the total thickness locally attaining 12 meters.

Southeast of the Tarlanat ravine, rocks of the Buruntau formation are represented by breccia-conglomerate, either massive or interbedded with rubbly gravel. Their thickness is inconsistent, because of their subsequent erosion. Where they form the base of alluvial fans (i.e., at the base of middle Pleistocene deposits), their thickness does not exceed 1.5 or 2.0 meters (Sarybulak and Shiyentas ravines). In outcrops, their thickness increases to 8 meters.

With regard to the origin of the Upper Gobi conglomerate, V.A. Obruchev correctly regards it as a sheet-flood formation [10] of a period when precipitation became periodical rather than regular, coming in downpours which resulted in sheet flows.

The exclusively gray color of lower Pleistocene alluvial and alluvial fan deposits suggest a climate more rigorous than in the Eopleistocene. Such a strongly continental climate naturally resulted in an intensive physical weathering of bare Chu-Ily Mountain rocks, unprotected by vegetation. After a quick spring thaw and occasional summer downpours, masses of rubble were rushed down the slopes, carried by mud flows, and to the foothills, obliterating the relief. Small lakes were formed as a result of damming, with loam and marl deposited in them.

However, this cooling was not great; otherwise, it would have brought about a sharp increase in humidity which, in turn, would have affected the nature of the deposits. A confirmation of that is the above-listed mammalian fauna from Pleistocene deposits of the neighboring regions of Kazakhstan, which would have existed in a temperate climate. Cold-loving forms are absent there.

6. Middle Pleistocene Q_2^2

a) Alluvial Deposits

(Shoshkaul'gen formation) Q_2^{2al}

The terrace described below is developed on both the right and the left slopes of the Chu valley. Its surface slopes to the west, i.e., down the stream; and to the north, toward the river channel.

In the western part of the area under study, it is made up of gray pebbly to pebble-free sands, locally overlain by gray loam or sandy loam, 0.3 to 1.0 meter thick. The total thickness of these deposits is not over 5 meters.

Their thickness increases to the east. Thus 20 kilometers south of the Maldabay well, Shoshkaul'gen yellow-gray sandy calcareous silt with lenses of gray siltstone are 24.5 meters thick; 40 kilometers south of there (outside the area under study), these rocks are 71 meters thick. These are yellow-gray, well-washed and sorted silts, underlain by gray pebbly sand, 25 meters thick.

To the southeast, in the Sarykobu and Shoshkaul'gen area, middle Pleistocene alluvial deposits are represented by coarse, well-washed and sorted silt, over 192 meters thick.

On the right bank of Chu River (in the area of Kok-Tobe settlement), the middle Pleistocene terrace is made up of gray pebbly sand, below, and of sandy loam above, a total of 12 to 24 meters.

It is interesting to trace the change in relative height of the surface of deposition of middle Pleistocene sediments. At Kok-Tobe, it lies 10 meters below the Chu water level; near Sarykobu and Shoshkaul'gen, it is more than 60 meters below it, with the terrace apparently not less than 40 kilometers wide.

Forty kilometers south of the Moldabay well, the basement of the middle Pleistocene deposits is 30 meters above the water edge; it gets lower, farther west, then rises again. It stands about 20 meters in the Lake Tuz-Kul' area and west of there; 30 meters, south of Dzideli; and finally 30 meters, south of Kosobazhon where the terrace is but 10 kilometers wide.

These data on the position of the base of middle Pleistocene deposits, their thickness and composition, suggest that intensive subsidence in the area of Sarykobu and Shoshkaul'gen, in middle Pleistocene, became, to the west, an area of relative uplifts; in other words, intensively differential tectonic movements took place in the Chuysk trough, at that time.

The great thickness of silt in the area of subsidence (over 192 meters) precludes their assignment to a floodplain facies, as understood by Ye.V. Shantser [14], because such a high flood plain is inconceivable. Silts represent the facies of a river split into a number of channels; it is a flood facies, close to the floodplain facies of Ye.V. Shantser but not identical with it. A flood facies is not contemporaneous with the underlying pebbly sands, but is younger.

In narrow stretches of the valley, there is the usual superposition of the floodplain facies over the channel, as a result of the channel's shifting.

Deposits of this terrace are not barren of

organic remains. In the area of communal farm Novyy Put', located on the right bank of the Chu, 30 kilometers above Kok-Tobe, i.e., outside our area, the middle Pleistocene terrace carries many remains of mollusk shells (identified by T.M. Mikulina), such as Eulota (Leucozonella) rubens (Mart.), and some Succinea putris (L.), both of which persist into the present.

Spore-pollen analysis suggests that the vegetation cover of the middle Pleistocene ancestral Chu River valley was similar to the present one. It can be stated, nevertheless, that the middle Pleistocene climate was more humid than the lower Pleistocene and the present. A proof of that is the nature of middle Pleistocene alluvium which contains a locally thick (25 meters) channel facies represented by pebbly sand. The latter must suggest a middle Pleistocene Chu flow considerably greater than the lower Pliocene one. The question arises, was this stronger flow connected with the humidity rise in Tyan'-Shan' as a result of the latter's uplifts, or was it due to a more humid climate on the Chu steppes?

A study of middle Pleistocene alluvial fan deposits at the foot of the Chu-Ili Mountains suggests that they, unlike those of the lower Pleistocene, are coarser, better sorted (still poorly sorted, on the whole), and better stratified. If we add the considerable incision of middle Pleistocene valleys into lower Pliocene rocks, it becomes clear that the contemporaneous streams in the Chu-Ili mountainous area became more powerful and longer lasting. That could have been achieved only by a generally more humid climate: the Chu-Ili Mountains could not have been collectors of moisture, because of their low elevation.

The more humid climate of the Chu steppes in the first half of the middle Pleistocene could have been an effect of the glaciation which set on in the mountains and in the north of western Siberia.

Evidence of the advent of a colder climate in the adjacent regions of Kazakhstan are the findings of early forms of Mammonteus primigenius, by M.I. Lomonovich, in the vicinity of Alma-Ata. This species was adapted to a cold climate and open spaces. Other forms, Felis sp., Camelus sp. (knoblochi?), Cervus (Megaceros) ruffi, C. sp., Equus sp., Paralephus trogontherii, collected by various investigators in different areas of Kazakhstan, also suggest the presence of open spaces there, at that time.

When dry conditions set in, in the second half of the middle Pleistocene, the Chu flow was reduced, and silt was deposited in the flood area. This sedimentation stage probably corresponded to an interglacial stage in the adjacent regions.

b) Alluvial Fan Deposits Q_2^{2pl}

Middle Pleistocene alluvial fan deposits are traceable along the western slope of the Chu-Ili Mountains, from Chatyrla Mountain in the northwest to the Khan-Tau in the southeast. They rest on rocks of the Andassay, Kenshagyr, and Buruntau formations, and occasionally on the Upper Cretaceous and Paleozoic. In the Kenshagyr ravine, at the Tortkul' mesa, they are represented chiefly by gray, yellow-gray, and brown, fine-grained argillaceous, pebbly sand, locally with fine, slightly rounded gravel, changing laterally to rubbly gravel, coarse sand, and marl.

There are small-pebble gravels lying at the very base of the pebbly sands; they are rubbly, gray, and 0.5 to 2.5 meters thick. The maximum overall thickness of middle Pleistocene rocks is 6.8 meters.

Going down the Kenshagyr Valley, these deposits are replaced by yellow-gray sand, clay, and loam; i.e., they become finer while their thickness ranges from 7 to 10.5 meters.

Ostracods occur in these clays and loams, 6 kilometers southwest of the Tortkul' mesa. G.F. Shneyder has identified the following, from our collection: Limnocythere manijtschensis Neg., Cyprideis littoralis Br., Trachyleberis pseudoconvexa (Liv.), Ilyocypris gibba (Ramdohr), Il. brady Sars.

According to G.F. Shneyder, these ostracods are Quaternary.

Middle Pleistocene deposits are almost completely missing in the Andassay ravine; they occur in lenses at the base of upper Pleistocene rocks, 2 kilometers south of the Andassay-Karaoba sopkas (gypsiferous sand with rubble, and loam). At the foot of the Mayzharlygan—Khan-Tau Mountains, middle Pleistocene formations are represented by a monotonous succession of gray boulder and rubbly gravel with a matrix of calcareous, argillaceous sand. Horizontal to cross-stratification has been locally observed in the gravels. Total thickness, 1 to 7 meters.

Middle Pleistocene rubbly gravels are present in those canyons where modern streams are practically inactive, depositing only fine clay. The inference is that the middle Pleistocene climate was more humid than the present.

7. Upper Pleistocene Q_2^3

a) Alluvial Deposits

(Saroy formation) Q_2^{3al}

Upper Pleistocene alluvial deposits form the first higher than flood-level terrace of Chu River, traceable in broad bands (locally up to 40 kilometers) on both slopes of the valley; they also fill up the semi-isolated Saroy depression, on the northwestern continuation of the first higher than floodplain terrace of the right bank.

Involved in the structure of this terrace are pebbly sand, gravel, and loam, making up two well-defined facies in narrower parts of the valley: the channel and the floodplain. They rest on Eopleistocene, Upper Oligocene, Middle Oligocene, Upper Cretaceous, and Paleozoic formations. In most places, this terrace is separated from the upper one by a distinct escarpment.

In the eastern part of the Saroy depression, upper Pleistocene deposits are represented by gravel changing laterally to pebbly sand, overlain by loam. The maximum total thickness of upper Pleistocene deposits here is 3.2 meters, usually 2.5 m. The ratio of the channel to the floodplain facies is about 1:1.

To the west, the channel facies of upper Pleistocene alluvium is represented chiefly by gray pebbly sand, locally with a slight addition of fine gravel (not over 8%), while the floodplain facies is represented by gray porous loam. The total thickness of this formation ranges from 4.5 to 10 meters. The ratio of the river facies to the floodplain facies is mostly 2:1, occasionally 5:1.

Sixty kilometers southeast of the Saroy depression, the structure of this formation is not clear for lack of mining works and boreholes. It has been uncovered only near the mouth of the Tarlanat ravine where it is thin (4 to 7 meters) and has a distinct twin layer structure. Interchanges of loam and gravel leave no doubt of this being the channel and floodplain facies of an upper Pleistocene ancestral Chu River.

A typical feature of the river facies, is the presence of rubble along with gravel. The rubble was brought in from the Mayzharylgan Mountains, by waters rushing down the canyons, when the upper Pleistocene ancestral Chu River was flowing in that vicinity. Thus the alluvial formation, here, was "contaminated" by alluvial fan deposits. The channel-floodplain deposits ratio is about 1:1. The structure of upper Pleistocene deposits changes abruptly, toward the central part of the depression: they grow thicker, with a changing ratio between the two facies.

About 60 to 70 kilometers southwest of the Chu-Ili Mountains, the thickness of upper Pleistocene deposits increases 12 to 14 times, reaching 85 or 100 meters, with the "alluvial stage" formed chiefly of silt from a flood facies. Such thicknesses, considerably greater than the usual for plainland river valleys, reflect upper Pleistocene subsidences in that area. These subsidences are confirmed also by the position of the alluvium deposition surface. Southwest of Gulyayevka village, it lies 50 meters lower than the Chu River channel; and it is 60 meters below it, southwest of Kokterek village. Northwest of Gulyayevka, down the Chu, the basement of the Saroy formation rises. Approximately 80 kilometers away, it is only 12 meters below the water edge, on the average.

South of Karatal, the base of the Saroy alluvium is above the river level, 14 meters on the average, while the ratio of the channel (pebbly sand) and the floodplain facies is about 1:1, reflecting a considerable decrease in the floodplain facies. The total thickness is 19 meters. This ratio, along with the small thickness, suggests an area of upper Pleistocene uplift, south of Karatal, relative to the eastern areas.

This rising area extended from Karatal, west to Kosobazhon, a distance of about 110 kilometers. The first higher than flood-level terrace, here, is always the basal one, with the surface usually standing 8 to 11 meters above the water edge. The formation thickness is not over 11 meters, usually 5 to 8 meters. It consists of fine- to medium-grained sand, in places with pebbles and occasional gravel, and gray calcareous sandstone and loam changing laterally to sandy loam. The latter two are not important in upper Pleistocene alluvium, as their thickness does not exceed 1 or 2 meters. Thus, river facies are paramount in the alluvial formation. West of Kosobazhon, the basement of the Saroy alluvium once more plunges below the water level, thereby marking a new zone of relative subsidence, in the upper Pleistocene.

The basement of the Saroy formation rises upstream from the areas of upper Pleistocene subsidence around Gulyayevka-Kokterek village. At Koskuduk village (near Chu station), it is 35 meters below the water level, and 25 meters below at Berlik settlement. Conspicuous in this formation are sandy facies, with upper Pleistocene deposits in lateral contact with the middle Pleistocene. Lower Pleistocene alluvium is missing because the Chu was located somewhat to the southwest, at that time. For that reason, N.N. Kostenko's 1950 classification of the Quaternary for this

area, according to which upper Pleistocene deposits are underlain by middle Pleistocene, and the latter by lower Pleistocene, is erroneous.

The first higher than floodplain terrace, both in this area and south of it (above Novo-Troitskoye village), carries a large amount of fresh water and terrestrial mollusk shells of species still living here.

In 1949, Z.I. Gur'yeva collected Unio sp. and Anodonta sp. in the Saroy depression. T.M. Mikulina identified the following species in our collection from different localities on this terrace: Eulota (Leucozonella) rubens (Mart.), Vallonia pulchella (Müll.), Cochlicopa lubrica (Müll.) var. nitens Callenstein, Succinea putris (L.), and Planorbis planorbis (L.).

The pollen in these sediments is mostly of non-woody associations, with the tree pollen accounting only for 0.3 to 1.6%. On the whole, a flora similar to the present one prevailed here during the upper Pleistocene. However, the nature of alluvial and alluvial fan upper Pleistocene deposits suggests a climate more humid than the present, in the latter half of the Pleistocene, and apparently differing but little from the climate of the preceding age. In grain size and degree of rounding and sorting, upper Pleistocene deposits, both alluvial and alluvial fan, hardly differ from those of the middle Pleistocene which either underlie them or are located at the same meridian. This means that upper Pleistocene streams were about as strong as those of the middle Pleistocene. At the same time, upper Pliocene sediments are much coarser than the Holocene above them or on the same meridian.

It is probable that upper Pleistocene climate in the Chu steppes was fairly rigorous. This is confirmed by the upper Pleistocene faunal assemblage from the adjacent regions of Eastern Kazakhstan, associated with different genetic types of deposits, including the "bald" terrace.

The paper of V.S. Bazhanov and N.N. Kostenko (see above) gives a list of mammals collected by various investigators in different parts of Eastern Kazakhstan: Bison priscus deminutus, Bos sp., Ovis ammon, Saiga imberis (-tatarica), Capreolus sp., Rangifer tarandus, Cervus cf. elaphus, Equus caballus, Rhinoceros antiquitatis, and Mammonteus primigenius. According to them, the remains of such a cold-loving animal as Mammonteus primigenius (late form) occur throughout eastern Kazakhstan, as far south as the line Aral'sk-Chu station-Alma-Ata. Inasmuch as Mammonteus primigenius and other forms are denizens of open spaces, it is to be assumed that there was no forest in eastern Kazakhstan, at that time, except perhaps for groves along rivers.

The cooler and more humid climate of the first half of the upper Pleistocene in the Chuysk steppes obviously was the effect of a new glaciation in the adjacent provinces. At that time, in the Gulyayevka subsidence zone, the Chu deposited pebbly sand whose thickness locally reached 23 meters.

In the second half of the upper Pleistocene, because of the advent of dry climate, the stream slackened considerably. It became split up into channels where silt was deposited.

Corresponding to the first higher than flood-level terrace of the Chuysk steppes is a terrace in the north Aral region, which A.L. Yanshin [16] designates as the second higher than floodplain terrace. He states, "Because of the peculiar lithology of this terrace, it differs in its vegetation cover as well. In most places, there is only sparse Anabasis salsa brush, with interspersed bare clay areas. Because of that, the second terrace, as seen from an airplane, stands out because of its light color" ([16], p. 645).

In the Aral-Turgay plain and the adjacent regions of western Kazakhstan, deposits of this terrace carry remains of the northern forest flora and fauna: oak leaf imprints, cones of spruce and larch, remains of beaver and peat elk [16].

A.L. Yanshin correctly relates the faunal features of this north Aral region to glaciation. Inasmuch as this terrace rests on the Khvalynsk transgression, according to him, its age must be Khvalynsk.

It should be noted that there is no contradiction in correlating the first terrace of the Chuysk steppes and the second north Aral region terrace. The difference is in nomenclature, only. It is obvious that by "the second higher than floodplain terrace" A.L. Yanshin means a terrace immediately above the so-called high floodplain, because this is the terrace nomenclature previously used by this author [3]. We believe it is more correct to designate the "bald" terrace of the north Aral region as the first higher than floodplain; or the second, if the flood plain is regarded as the first terrace. It also is obvious that some confusion in the terrace nomenclature is due to the inadequate understanding of the formational process of a floodplain, with a "high" and a "low" floodplain usually recognized, with different stages in both. Thence, the different counting of terraces. Ye.V. Shantser [14] has demonstrated convincingly that the formation of floodplains

is very complex and that the presence of escarpments in them does not necessarily mean a new erosional cycle. It appears that a differentiation of "high" and "low" flood-plains in western Kazakhstan is not adequately substantiated by field data. Indeed, they probably represent a single flood plain terrace formed as a result of a single erosional cycle.

Correlative with this terrace, in the lower Syr-Dar'ya course, is the Dar'yalyk-Takyr plain (also marked by the light coloring of its surface), which N.N. Kostenko [5] erroneously believes to be middle Pleistocene.

Syr-Dar'ya alluvial deposits older than the Dar'yalyk-Takyr are developed south of the latter.

b) Alluvial Fan Deposits Q_2^{3pl}

Upper Pleistocene alluvial fan deposits are developed in the same areas with those of the middle Pleistocene, i.e., along the southwestern slope of the Chu-Ili Mountains. They rest with an erosional unconformity on the middle Pleistocene, Miocene, and Paleozoic.

East of Tortkul' mesa, upper Pleistocene alluvial fan deposits are represented by poorly rounded gray gravel, with a maximum thickness of 8.7 meters, with lenses of coarse-grained and to a smaller extent fine-grained sand. As exposed in the mine drift walls, the gravel beds are cross-laminated. Higher up the ravine, they become thinner, down to one meter; going down, they are replaced by pebbly sand with gravel, changing to pebbly sand, 1 to 2 meters thick. A similar regular replacement of gravel by sand, down the ravine, with a parallel increase in the thickness, has been observed in the Andassay ravine. At the mouth of the latter, at the Saroy depression, these deposits are represented by yellow-gray, fine-grained argillaceous sand with well-rounded small pebbles; their thickness is 2 to 2.5 meters.

To the southeast, in the Tarlanat and Shiyentas ravines, upper Pleistocene alluvial fan deposits, 1 to 4 meters thick, consist of gravels (35 to 45%), with small pebbles and very coarse sand, and with a gray, coarse-grained sand for matrix. Southeast of the Khan-Tau Mountains, these deposits are rubbly gravel (50%), with lenses of sand and conglomerate, and which chunks and boulders, up to 25 centimeters, which have originated in the disintegration of the mountains. Thickness, 2 to 6 meters.

The alluvial fan deposits, which we assign to the upper Pleistocene, carry floral and faunal remains. Thus, according to Ye.D. Polyakova (1952), remains of Equus asinus sp. were identified by I.M. Gromov from a 6 meter terrace along a northern tributary of the Sarybulak. The same deposits contain fresh-water mollusks which, according to I.V. Danilovskiy, persist into the present.

This discovery has made Ye.D. Polyakova date these sediments as recent. This is not correct, because the most recent sediments are located 6 meters deeper, in that area. According to Ye.D. Polyakova, these and similar deposits along a right tributary of the Tarlanat, carry pollen of Chenopodiaceae, Cruciferae, Juniperous, and Ericaceae, along with mushroom spores and vegetable tissue. According to V.V. Zauyer, the sediments with these vegetable remains were formed in a fairly humid climate, as witness the sizable amount of mushrooms and woody plant tissue, although the vegetation was on the whole of a steppe character. These data (together with the nature of the faunal assemblage and the lithologic features of upper Pleistocene alluvium) also confirm the prevalence of comparatively humid upper Pleistocene conditions.

8. Holocene Q_3

a) Alluvial Deposits

(Chuysk formation) Q_3^{al}

Holocene alluvial deposits are involved in the structure of the Chu Valley floodplain. The latter is mostly marshy and overgrown by reeds and meadow grasses. Its width varies greatly, from 1 to 5 kilometers in narrower stretches to 25 or 30 kilometers in flood areas.

Like the older alluvium, the floodplain exhibits two levels: the lower, a channel level; and the upper, a blanket representing a floodplain facies. Stagnant alluvium is present in many ox-bow lakes below Gulyayevka settlement. This alluvium is represented by a dark-gray to almost black, humus-rich substance.

In most places, the channel facies of Holocene alluvium consists of gray to dark-gray fine-grained, humus-carrying sand, locally saline. Much less common are gray pebbly sands with gravel, occurring in narrower parts of the Chu Valley. Resting on them is a blanket of gray to dark-gray humus-carrying, saline loam changing laterally to reed peat. The boundary between the two facies is not sharp, and the alluvial formation, on the whole, looks like a floodplain facies.

Inasmuch as the Chu River is broken up in the summer into isolated stretches of shallow

water, subject to intensive silting up, the difference between fairway and bank shoals is obliterated. In addition, in many places the bank shoal facies is represented by fine-grained sand changing to the mid-stream facies of dark gray oozes with sand. Thus, because of these peculiar river conditions, a sedimentary sequence originated, the reverse of that prevailing in plainland rivers of a temperate climate.

A study of Holocene alluvial facies shows that flood-facies predominate southeast of Karatal, i.e., in the Gulyayevka zone of subsidence, where the floodplain is not over 1 or 1.5 meters high. West of there, channel facies predominate in the floodplain sections (although represented by fine-grained sand), and the floodplain is two or three times higher. By analogy with older deposits, this is due to a different direction of tectonic movements: a subsidence in the Gulyayevka area as against the uplifts west of there.

It is characteristic that the uninterrupted flow of the present Chu River is traceable only as far as Gulyayevak (Furmanovka). In flood, the stream barely reaches Lake Bol'shoy Kamkalykul'. The lowest point affected by highest flood waters is Lake Ashi-Kul'. West of that lake, the Chu channel is perfectly dry; the upper Pleistocene ancestral Chu, judging from the development and structure of the first higher than floodplain terrace, from Gulyayevka to Lake Ashi-Kul' and farther west, had a permanent and fairly strong flow. The only factor which affected the hydrogeologic status of the upper Pleistocene ancestral Chu River was the advent of a drier climate.

b) Alluvial Fan Deposits Q_3^{pl}

Alluvial fan deposits of the present age fill up the valleys of intermittent streams flowing down the Chu-Ili Mountains and the Kazakh highlands, and into the Chuysk trough. However, the deposits of modern streams, unlike those more ancient, such as the upper Pleistocene, do not form a continuous train along the foot of the mountains but rather are localized in valleys. For that reason, they perhaps should be regarded as spoon-shaped alluvium rather than alluvial fans.

In the lower courses of the Andassay and Kenshagyr ravines, Holocene deposits are represented by loam, 1 to 2 meters thick, replaced by about the same thickness of rubbly alluvium at the foot of the Chagyrla Mountain. It may be inferred, therefore, that their spring-time flows were small, effective only in the mountains. Southwest of the mountains, they spread wide, to form very small lakes depositing clayey material. It is known that comparatively thick gravels were deposited here, during the upper Pleistocene.

The sharp reduction in the modern flow is undoubtedly related to the drier climate; this is also confirmed by the lack of correspondence between the width of these valleys and the volume of their flow. Thus the Kenshagyr valley, in the Tortkul' mesa area, attains a width of four kilometers — without any flow, even during the spring thaws. It follows that such a valley was developed in a more humid upper Pleistocene climate.

Present deposits in ravines to the southeast are represented chiefly by rubbly gravel, in their highland courses; some distance to the southwest, away from the disintegration zone, the gravel gives way to loam and sandy loam, with poorly rounded pebbles and rubble toward the base.

The thickness of deposits varies greatly in any given ravine, with a maximum up to 5 meters somewhat below the source; they grow thinner, higher up, locally only 10 to 30 centimeters thick. In places, they are altogether lacking, exposing the rocks at the bottom of the ravine.

CONCLUSION

The structure of Quaternary alluvial formations in the Chuysk steppes confirms on the whole the Ye. V. Shantser classification as regards the relationship of the channel and floodplain facies, in narrower parts of a valley. In the subsidence areas, on the other hand, we have identified flood facies not correlative with his floodplain facies [14]. Under such conditions, the flood facies should be regarded not as having originated as a result of channel migration, i.e., contemporaneous with the underlying gravel, but are older than the latter.

A study of alluvial formations in the Gulyayevka subsidence zone suggests that this area underwent two glaciations — a middle Pleistocene and an upper Pleistocene — when the climate became substantially cooler and more humid, and the ancestral Chu River became swollen.

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Geological Institute
of the Academy of Sciences, U.S.S.R.
Moscow

Received March 18, 1959

GLAUCONITE FROM PALEOGENE DEPOSITS OF THE STALINGRAD VOLGA REGION[1]

by

G. I. Nosov and F. Ye. Bogokina

Glauconite is one of the rock-forming minerals in Paleogene formations of the Stalingrad Volga region. The Paleogene is represented here by marine, green-gray, terrigenous deposits: sand, friable sandstone, and siltstone, interbedded with silty claystone and clay. Lithologically, the sequence is differentiated into three formations, reading upward: the Proleysk, Tsaritsin, and Buchak. Its overall thickness is over 100 meters.

A typical feature of Paleogene deposits is the abundance of glauconite. Sandy-silty facies of these rocks consist mostly of quartz (20 to 80%), glauconite (10 to 70%), and an addition of feldspar (4 to 10%), chalcedony (1 to 4%), pyrite (1 to 3%), and accessory minerals (up to 1%).

This study on glauconite was done by various methods: microscopic, chemical, spectroscopic, thermal, X-ray, electronographic, and electronmicroscopic. Glauconite in rocks was studied in thin sections under the microscope; for all other analyses, it was first isolated by electromagnetic separation from friable rocks. The dispersion of grains was done ultrasonically.

It has been determined, as a result, that glauconite from all three Paleogene formations is not substantially different; accordingly, its description, as given below, pertains to the entire Paleogene of the Stalingrad Volga region.

GENETIC VARIETIES OF GLAUCONITE

Paleogene deposits of the Stalingrad Volga region carry glauconite of three genetic varieties: round to granular, foliated, and pigmentary (Fig. 1). The first one is the most common; the other two are fairly rare.

1. Round to granular glauconite usually occurs in bean-like to reniform grains peculiar to concretionary-gel formations. The grain size ranges from 0.1 to 0.5 millimeter, being usually 0.2 to 0.3 mm.; their color is dark green, when fresh, less commonly greenish black; the surface is rough and dull to smoothly polished in dark grains.

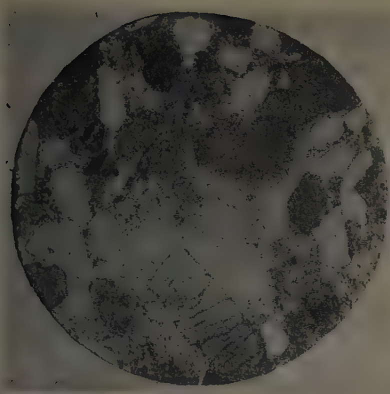
As seen in thin section, the grains are lobate, less commonly irregularly elliptical, occasionally approaching circular (Fig. 2). Their green color is of a different intensity, occasionally becoming yellowish green. Typically, the color intensity is uneven throughout a section. This feature is peculiar to glauconite in general, and has been studied in detail by A.V. Kazakov [5]. In his opinion, it is connected with the uneven pigmentation of grains by the chromophore substance.

The presence of fine inclusions of pyrite (commonly limonitic), phosphate, carbonates, etc., is typical of glauconite grains.

Highly magnified (over 300 times) thin sections reveal a micro-aggregate structure of rounded glauconite grains, consisting of closely packed and haphazardly oriented scales measuring about one micron. With the Nicol prisms crossed, they produce a typical micro-aggregate polarization of the grains, in greenish-gray hues. This aggregate micro-granular structure is evidently related to the primary gel nature of glauconite. This is suggested by the presence of a fine (0.01 to 0.03 millimeter) outer shell, in many grains, fairly easily separable from the main body by mechanical action. Seen in thin section, this shell is commonly peeled off and the space between it and the kernel, up to 0.05 millimeter wide, is filled some times by a secondary opal substance (Fig. 3). The peeling off of the shell is related to the aging of the kernel.

The refractive index for round-grained glauconite, as measured in immersion fluids under the microscope, ranges from 1.588 in light-colored grains to 1.601 in the dark ones;

[1] Glaukonit paleogenovykh otlozheniy Stalingradskogo povolzh'ya.



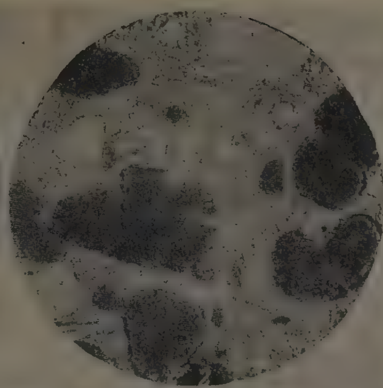
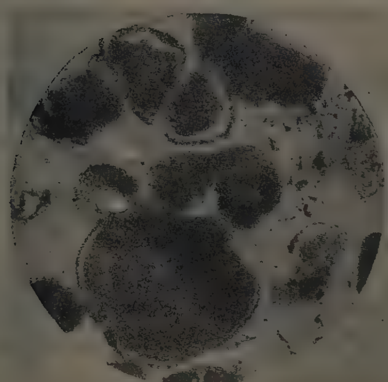


FIGURE 1. Round to granular, foliated, and pigmentary glauconite in siltstone. Thin section; magnification 64X; without analyser.

FIGURE 2. Lobate and oval outlines of bean-like and reniform round-grained glauconite in argillaceous sandstone. Thin section; magnification 64X; without analyser.

this is determined by their iron content. Grains discolored by hydrochloric acid have a lower refractive index, down to 1.470, approaching that for opal. This is fully consistent with the siliceous composition of the skeleton.

In order to obtain more precise refractive indices for round-grained glauconite, oriented aggregates were prepared from a dried-up film of ultrasonically dispersed grains. The following indices were determined in an immersion medium: $\gamma' = 1.622$ and $\alpha' = 1.602$, whence birefringence $\gamma' - \alpha' = 0.020$. It is significant that grass-green, oriented aggregates of glauconite are fan-shaped, sheaf-like,



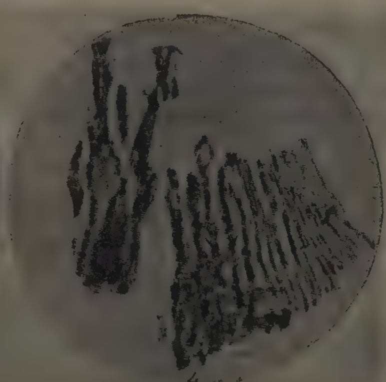


FIGURE 3. Peeling-off shells (aging of gel) on round-grained glauconite from friable sandstone. Thin section; magnification 64X; without analyser.

FIGURE 4. Fan-shaped and sheaf-like aggregates of oriented glauconite particles. Immersion preparate; magnification 340X; without analyser.

and occasionally spiral, which is typical of the montmorillonite group minerals (Fig. 4). It is possible that the origin of such aggregates is determined by the nature of cations adsorbed, and by a finer dispersion of the substance under study.

The origin of glauconite is fairly well known, at the present time; it is discussed in many papers [1, 2, 5, 6, 7, 9, 10, 11].

The authigenic origin of round-grained glauconite in Paleogene deposits is beyond doubt. This is confirmed by its occurrence in typically marine, relatively shallow-water deposits, where the predominant terrigenous fraction of sediments is considerably finer than the glauconite grains. Its grains were formed apparently in an originally deposited silty-clay ooze, during its diagenesis in a slightly oxidizing to a neutral medium (for Eh), in the upper sedimentary oxidizing zone. Processes of redistribution of silica, iron, and aluminum, in solution in the bottom seawater layer of normal salinity, and in the ooze solutions, were active in this potassium-rich environment, with resultant coagulation of an alumino-ferro-silicate gel into round-grained glauconite.

Rare findings of angular glauconite grains can hardly be indicative of their terrigenous origin. Such grains are most likely a result of mechanical action of bottom currents reworking the bottom sediments.

2. Foliated glauconite occurs in typical, elongated platy grains with uneven ends, 0.4 to 0.6 millimeter long and 0.1 to 0.2 millimeter wide. The grains are green to light green, with a dull surface. As measured in thin section, this variety accounts for about 10% of all glauconite in these rocks.

As studied in thin section, the glauconite plates show evidence of cleavage, being made up of elongated, parallel scaly microlites, various shades of yellow-green in color. Pleochroism is clean-cut, from dark green to pale yellow; a direct extinction with reference to the cleavage planes; a high birefringence, bringing about fairly vivid interference colores. The refractive indices, as measured in immersion fluids are as follows: $\gamma' = 1.591$ to 1.620; $\alpha' = 1.571$ to 1.579; $\gamma' - \alpha' = 0.020$ to 0.023; positive elongation of prismatic grains; the mineral is biaxial and negative.

Some grains contain residual greenish-brown, pleochroic segments of a biotite-type micaceous mineral, which suggests the origin of foliated glauconite by metasomatic alteration of clastic micas, during diagenesis processes.

3. Pigmentary glauconite has been visually determined only in silty argillaceous rocks, from greenish spots of various intensity. However, a microscopic study of thin sections reveals the presence of pigmentary glauconite in silty arenaceous rocks, as well, although to a considerably smaller extent, also as a component of the finely-dispersed opal-argillaceous cement.

It is possible to establish from a study of thin sections, under high magnification, that pigmentary glauconite occurs in the rock in fine (<0.005 millimeter) isometric grains, green, and slightly pleochroic. These scales fill up the interstices between clay minerals of the montmorillonite and hydromica type, which form, along with globular opal, the cementing groundmass. Pigmentary glauconite has been observed in places filling small cracks and isolated pores in the rocks.

A more detailed study of pigmentary glauconite, in thin section, was unsuccessful because of the small size of the particles; an application of other and more precise analytic methods is hampered by the impossibility of its isolation in a pure state.

The nature of the distribution of pigmentary glauconite in rocks, its secondary entry into the cracks, and its morphologic features, suggest a diagenetic origin of this mineral.

SOME PHYSICAL PROPERTIES OF GLAUCONITE

The standard hardness of glauconite is about 2.5, never exceeding 3; a green streak; grains are brittle, falling apart under pressure into a greenish powdery mass.

According to I.V. Kolomenskiy (Moscow Exploration Institute), the specific weight of Paleogene glauconite is 2.78 to 2.95; the volume weight, 1.11 to 1.35 g/cm^3. The high specific weight is related to its high iron content, while the low volume weight is due to a high porosity of the mineral, 54.2 to 62.4%. the latter also is the cause of its great swelling, hydrophilic nature, and the degree of compressibility under pressure.

This high porosity and the swelling capacity appears to relate glauconite to clays. However, for its high upper plastic limit (about 44), glauconite has a very depressed lower limit, which renders it practically non-plastic and radically different from clay minerals.

Glauconite is partially soluble in concentrated acids, and fully soluble in alkalies. When heated in concentrated hydrochloric acid, its grains become discolored, rough, dull, and porous, while still maintaining their form. Residual grains of discolored glauconite

are fully soluble in a 25% solution of caustic soda.

CHEMICAL COMPOSITION OF GLAUCONITE

The published results of the study of glauconite reveal a considerable variation in its overall chemical composition, depending on the facial features of its conditions of formation [5]. Such an inconsistency in composition hampers the compilation of its formula which is written differently by different students. For the same reason, different authors assign glauconite to different mineral groups.

According to the present concept, glauconite is a hydromica with aluminum replaced to a considerable extent by iron and magnesium [3].

The results of chemical analyses by M.V. Simonova (Gidroproekt) show that Paleogene glauconite is enriched in ferric iron whose average content is about 20%. Calcium is present in a small amount (0.64%), apparently being partially bound to phosphorus in a phosphate, and partially as a carbonate. The average molecular ratio of oxides of iron and aluminum is 1:2, which makes it possible to regard this glauconite as ferruginous. The silicate modulus reaches 3.23, on the average; this is in full accord with L.I. Gorbunova's data [2] and confirms the hydromicaceous nature of glauconite (Table 1). On the whole, the chemical composition of Paleogene glauconite establishes it as a hydrous alumoferrosilicate of potassium and magnesium, with ferric iron in ascendency over the ferrous.

These data from the chemical study are in agreement with those obtained by spectrographic analysis, by V.I. Malinina (VNIGNI, All-Union Scientific Research Institute of Petroleum Geology. According to her, Paleogene glauconite contains, besides the chemically identified elements, V, 0.06%; Ti, 0.05%; Ni, 0.01%; Zn, <0.01%; Li <0.01%; Mn, 0.008%; and Cu <0.001%.

Accordingly, on the basis of its qualitative content of principal elements, glauconite belongs to the hydromica group; its distinctive feature is its higher vanadium and titanium content.

THERMAL CURVES FOR GLAUCONITE

A thermal study was carried out by Ye. A. Shurygina (Soil Institute, Academy of Sciences, U.S.S.R.), on three samples of glauconite separated magnetically out of sandy rocks of the Proleysk, Tsaritsyn, and Buchak formations.

Table 1

Chemical Composition of Pure Glauconite From Paleogene Deposits

Formations	Content of Oxides, %														Molecular ratio	
	SiO_2		Fe_2O_3	FeO	Al_2O_3	MgO	CaO	K_2O	Na_2O	TiO_2	SO_3	R_2O_5	Loss on ignition	Total	$\frac{Fe_2O_3}{Al_2O_3}$	$\frac{SiO_2}{R_2O_3}$
	Dound	Free														
Proleysk	46,38	5,22	20,37	1,64	10,23	2,61	0,68	6,36	0,32	None	0,14	0,26	5,76	99,97	1,27	3,27
Tsaritsyn	45,42	7,16	18,89	2:04	11,48	2,27	0,34	7,11	0,17	traces	0,03	0,16	5,19	100,26	1,05	3,26
Buchak	46,03	6,13	20,08	1,98	10,09	3,12	0,91	5,82	0,45	0,01	0,12	0,11	5,26	100,11	1,28	3,30
Average	45,94	6,17	19,78	1,88	10,60	2,66	0,64	6,43	0,31	traces	0,09	0,18	5,41	100,09	1,20	3,28

NOTE: Comma represents decimal point.

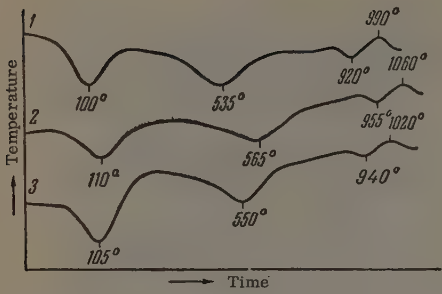


FIGURE 1. Round to granular, foliated, and pigmentary glauconite in siltstone.

Thin section; magnification 64X; without analyser.

As shown in Figure 5, the thermal curves for all three age varieties of glauconite are nearly identical, approaching those obtained by L.I. Gorbunova for Mesozoic glauconite [2] and by A.V. Kazakov for the Cretaceous [5].

Thermal curves for Paleogene glauconite show three well-expressed endothermal effects, with maxima at 100° to 110°C, and 920° to 955°C. The first endothermal effect, reflecting the loss of adsorbed water, is the most intensive, possibly because of the high degree of dispersion attained with ultrasonic equipment. The second endothermal effect, related to the loss of water of constitution, is located in a temperature interval typical of hydromicas in general. The third effect, coincident with the total collapse of glauconitic structure, is the weakest of the three. An exothermal effect follows the last endothermic one; it is rather poorly expressed and it reflects the recrystallization of amorphous products of glauconite disintegration.

The thermal curves for Paleogene glauconite are very similar to those for hydromicas, which confirms its hydromicaceous nature.

STRUCTURAL FEATURES OF GLAUCONITE

A study of structural features of glauconite was done in the VNIGNI laboratory (by Yu.M. Korolev and B.A. Anurov) by the X-ray and electronographic methods, on an isolated pure sample. The results of the X-ray analysis are given in Table 2.

The table shows that this glauconite exhibits, besides the lines common to clay minerals, an interplanar spacing typical of hydromicas; their diffraction pattern suggests that Paleogene glauconite belongs to the dioctohedral ferruginous variety of hydromicas.

Parameters of an elementary cell of glauconite, as determined by the two methods, turned out to be identical and very close to the corresponding values which B.B. Zvyagin gives for glauconite [4]. These results are given in Table 3.

Table 3 shows that, in its parameters, the elementary cell of this glauconite corresponds to the single-layer hydromica structural type. The somewhat higher value of parameter c probably is determined by a higher hydration of this mineral.

Table 3

Parameters of elementary cells for Paleogene glauconite, determined by two methods (in kX)

Method	a	b	c	β
X-ray	5.25	9.04	11.0	101°30'
Electronographic	5.20	9.02	10.3	101°06'

Table 2

Interplanar Spacing in Glauconite

(d is expressed in kx; I is the line intensity on the decimal scale; p is the diffraction effect in lines)

Line number	*d*	*I*	Line number	*d*	*I*	Line number	*d*	*I*	Line number	*d*	*I*
1	10.7	9	5	3,09	3*p*	9	2,122	1*p*	13	1,507	8
2	4,52	9	6	2,57	10	10	1,981	2*p*	14	1,3026	2
3	3,70	5*p*	7	2,405	7*p*	11	1,705	5*p*	15	1,2755	4
4	3,32	6*p*	8	2.259	2*p*	12	1,648	6	16	1,2554	4

NOTE: Comma represents decimal point.

ELECTRON-MICROSCOPIC DESCRIPTION OF GLAUCONITE

D. D. Kotel'nikov (Geological Institute, Academy of Sciences of the U.S.S.R.) studied finely dispersed particles of Paleogene glauconite, obtained by dispersion in an ultrasonic device, with the electron microscopic.

Figure 6 shows that the morphology of glauconite particles is generally close to that of hydromicas, with some specific features of its own. The particles are scaly, of various thickness, semi-transparent to opaque.

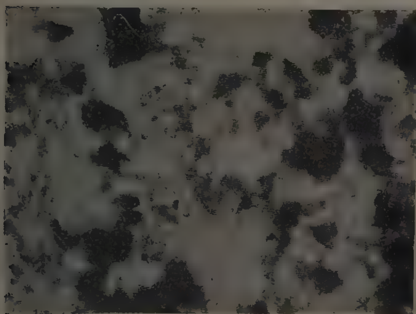


FIGURE 6. Morphology of ultrasonically finely-dispersed (< 1 micron) glauconite particles. Electron photomicrograph; 12,500X.

The scales are somewhat elongated, commonly isometric. The outlines are fairly sharp, uneven, cleaved in a step-like pattern, in places somewhat vague. In addition, there are specks of what probably are particles of opal which constituted the siliceous skeleton of round-grained glauconite.

WEATHERING OF GLAUCONITE

Glauconite is fairly easily affected by chemical and physical weathering, under hypergene conditions. Secondary weathering alterations in glauconite are especially noticeable in thin section. Glauconite grains from weathered outcrops are almost always corroded on the surface, often torn apart or else criss-crossed by numerous deep cracks. In the aeration zone, with its oxidation conditions, glauconite is most intensively decomposed. Leaching and oxidation of glauconite lead to a total loss of alkalies, a partial loss of silica and iron, and a change of lower oxides to hydroxides.

Weathering processes are especially vigorous in poorly-cemented rocks. Here, glauconite is intensively corroded and becomes greenish-yellow to drab-colored, because of the formation of iron hydroxides. As a rule, the peripheral part of such grains is considerably discolored, making a sort of fringe gradually changing to the darker kernels (Fig. 7).

Three stages of grain alteration are suggested in the weathering of glauconite. At the first stage, the surface of grains is corroded intensively, without any appreciable change in composition. The second stage is marked by a light color of the corroded peripheral part, with the kernel remaining intact. At the third stage, a nearly complete leaching of the green glauconite substance takes place, making the grains lose their characteristic color and rendering them porous.

The substance of such leached-out grains is silica, in the form of opal partly altered to chalcedony. The silica appears to make up a residual skeleton of glauconite grains whose pores still preserve greenish to brownish specks of the original mineral.

A more complex weathering process is revealed at times, determined by hydration of ferric iron in glauconite grains, without its complete leaching. Two fringes are formed in such grains, at the second weathering stage: a peripheral one, light-colored, consisting of opal; and a brown one, saturated with iron hydroxide. The third stage involves oxidation and hydration of the central part of such grains, rendering them brown and making them appear to merge with the second fringe.

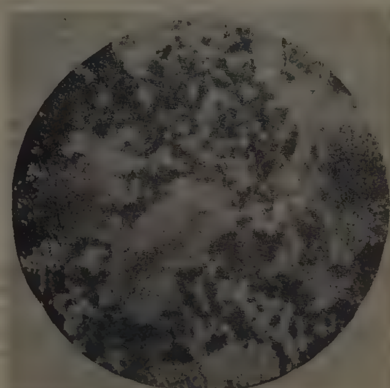


FIGURE 7. Peripherally corroded and discolored glauconite grains from a sandy-silt rock in the surface weathering zone. Thin section; 64X; without analyser.

There are isolated, corroded, brown grains of weathered glauconite with the residual silica skeleton fully replaced by iron hydroxide. It is possible that this limonitization of the residual skeleton was brought about by a secondary addition of iron hydroxide at later stages of glauconite weathering in the aeration zone. Also typical is the formation of opal and occasional limonite veinlets in weathered reniform glauconite grains. Such grains are commonly torn apart along the fissures, into reniform segments saturated with iron hydroxide and fringed with opal.

All these data only partly reveal the nature of glauconite weathering. This process undoubtedly is very complex and calls for further observation and study.

* * *

In conclusion, it can be stated that glauconite from Paleogene deposits of the Stalingrad Volga region, in its main features, composition, and morphology is very similar to glauconite from the Tertiary of the Russian platform and adjacent provinces [4, 7, 8].

Mesozoic glauconite, studied in detail by L.I. Gorbunova [2], from the Moscow basin Upper Jurassic and Cretaceous sections, also resembles Paleogene glauconites. This is especially true of glauconite from a sandy facies of the upper part of the shelf, which L.I. Gorbunova assigned to type one. This glauconite, like the Paleogene, is marked by the considerable size of its grains, by a high specific weight (2.7 to 2.9), a fairly high refractive index (1.59), a higher ferric iron content (19 to 20% Fe_2O_3) and potassium (6 to 7% K_2O), and a comparatively low free silica content (5 to 7% SiO_2).

Such a similarity in composition and properties of glauconite from several non-contemporaneous deposits, suggests that its origin is strictly related to a definite chemical environment wherein marine and terrigenous glauconitic formations are deposited [10].

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The All-Union Planning-Exploration Scientific Research Institute (Gidroproekt), Ministry of Electric Stations of the U.S.S.R. Moscow

Received December 17, 1958

THE KARSAKPAY ALKALINE AND NEPHELINE SYENITE MASSIF AND ITS STRUCTURAL POSITION[1]

by

T. G. Pavlova

This paper describes alkaline and nepheline syenites and notes the importance of albitization in the formation of these rocks. The structural position of the alkaline massif and the genetic relation of alkaline rocks with granites are considered.

* * * * *

GEOLOGIC POSITION OF THE MASSIF

The Karsakpay massif of alkaline and nepheline syenites, although areally small, is of interest as a definite genetic type of alkaline rocks related to a granitoid magma. A detailed study of this massif, under the guidance of N. A. Shtreys, has enabled the author to determine its structural position in the tectonics of the region, as well as some interesting features of origin of the syenites.

This massif is located in the western part of central Kazakhstan, 17 kilometers northwest of Karsakpay settlement. It was discovered and described by Ye. L. Butakova [3] in 1935, during her joint work with V. S. Sobolev in the Ulutau-Dzhezkazgan region of Kazakhstan. From 1950 to 1952, alkaline and nepheline syenites were mapped by K. A. Rachkovskaya. A microscopic study was done in the petrographic laboratory of the Karaganda Geologic Administration. The main conclusions of Ye. L. Butakova and K. A. Rachkovskaya are discussed in this paper.

The Karsakpay massif lies in the axial part of the Precambrian Maytyubin anticlinorium (Figure 1). The anticlinorium consists of several anticlinal and synclinal zones with a general sub-meridional north-northwest trend, within which individual brachymorphic folds are arranged en echelon. The central part of the anticlinorium is made up of lower Proterozoic rocks represented in anticlinal folds by the lower formation of porphyroid and to a smaller extent by quartz-muscovite schist with a few quartzite units. The lower formation is nonconformably overlain by a formation of quartzite, quartz-amphibole to feldspar-amphibole schist, and marble lenses. In synclinal folds, this section consists chiefly of rocks with colored minerals; in anticlinal folds, it consists of leucocratic gneiss (arkosic quartzite). Developed along the periphery of the anticlinorium, there is the third lower Proterozoic formation of acid volcanics.

The limbs of this anticlinorium are formed by assorted upper Proterozoic volcanics and by Cambrian deposits. Lying at the base of the upper Proterozoic are conglomerates carrying granitoid gneiss pebbles. The anticlinorium is complicated by a series of faults, trending meridionally, latitudinally, and to the northwest. The faults are of different ages, with the meridional being the oldest.

Precambrian granitoid gneisses and granites are involved in the structure of this anticlinorium, along with granodiorite and diorite. Syenite occurs in the middle of an anticlinal fold in the central part of the anticlinorium, which is made up of microgneiss and crystalline schist of the middle formation. This fold is confined to a triangular block with sides 13 to 17 kilometers long, defined by sublatitudinal folds in the south; trending northwest, in the southeast; and meridional, in the west. The western part of the structure is overlain by Carboniferous deposits. Small post-intrusion shears are present in the massif and in overlying rocks of this fold.

In plan, this massif with an area of 7 to 8 km^2 (Figure 2) is oval, somewhat elongated north-northwest, parallel to the anticlinal axis. The limbs of this fold dip to the east and northeast, at 40 to 55° in the east limb and

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[1]Karsakpayskiy massiv shchelochnykh i nefelinovykh siyenitov i yego polozheniye v tektonicheskoy strukture.

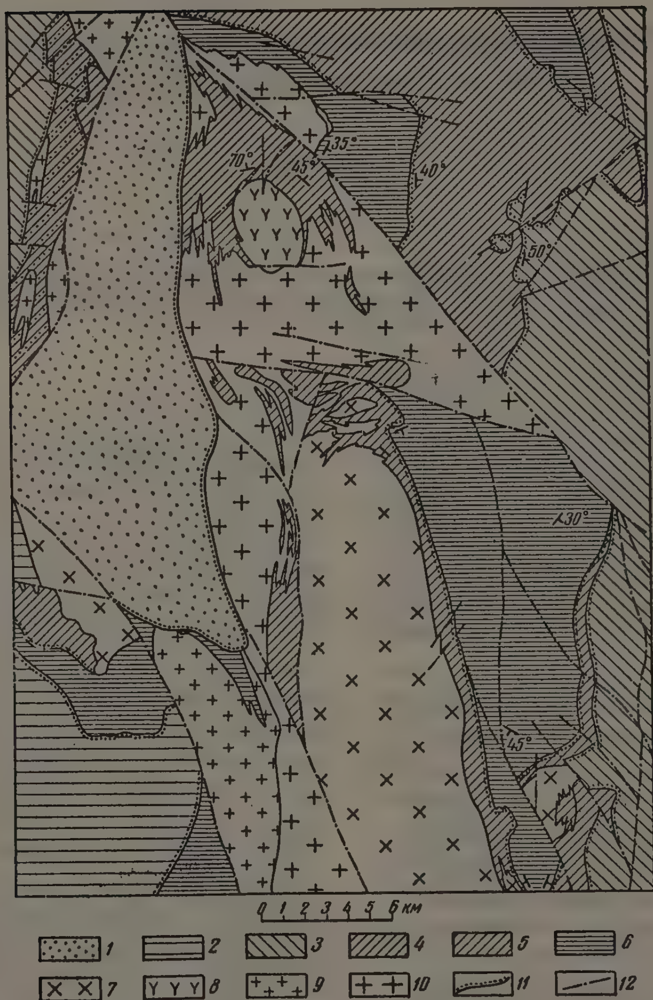

FIGURE 1. Geologic map of central part of the Maytyubin anticlinorium.

1 -- Middle Paleozoic sandstone and limestone; 2 -- Lower Paleozoic terrigenous clastic and other siliceous rocks; 3 -- Upper Proterozoic acid and basic extrusives; 4 -- upper formation of acid lower Proterozoic extrusives; 5 -- middle formation of quartzite and quartz-amphibole to feldspar-amphibole schist; 6 -- lower formation of porphyroids and quartz-muscovite schist; 7 -- diorite; 8 -- syenite; 9 -- granite; 10 -- granitoid gneiss; 11 -- transgressive contacts; 12 -- faults.

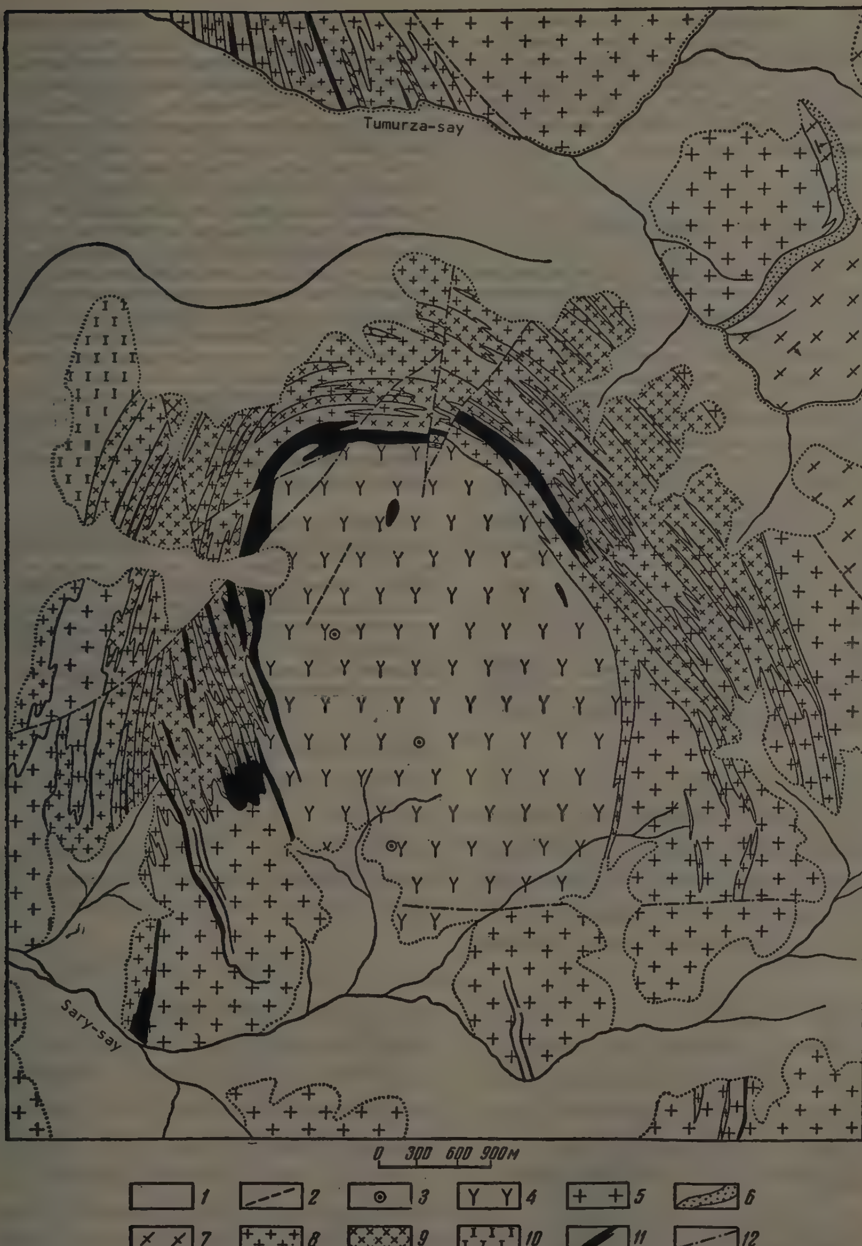

FIGURE 2. Geologic map of the Karsakpay syenite massif.

1 -- Unconsolidated deposits; 2 -- a dike of hornblende syenite porphyry; 3 -- nepheline syenite; 4 -- leucocratic and hastingsite alkaline amphiboles; 5 -- granitoid gneiss; 6 -- quartzite units; 7 -- porphyroids; 8 -- leucocratic microgneiss, aplitic gneiss; also fine-grained granitoid gneiss; 9 -- mesocratic microgneiss, dioritic gneiss, and augen gneiss; 10 -- quartz-diopside and quartz-tremolite rocks; 11 -- stratified bodies of amphibolites; 12 -- fault traces.

65 to 75° in the west limb. The northern closure is also steep; the southern closure has not been determined because the gneisses change to granitoid gneisses, in that direction. The contact surfaces of the massif are almost everywhere conformable with the strike of the enclosing rocks.

THE ENCLOSING ROCKS

The anticlinal fold, with which alkaline and nepheline syenites are associated, is made up of lower Proterozoic metasediments, whose variety in this area is explained by the sharp primary facies changes, as well as by their different degree of recrystallization. The least altered rocks are leucocratic quartz-feldspar to mesocratic feldspar-hornblende microgneisses — all massively banded, fine-grained rocks. The change from leucocratic, almost white, pinkish to light-gray gneisses and to mesocratic green-gray to brownish varieties is effected by a change in the ratio of quartz, feldspar, and hornblende. Mesocratic microgneiss exhibits a banded structure brought about by an orientation of the colored component. Microgneisses of various types have a microgranoblastic structure. Developed to a smaller extent are quartz-diopside and quartz-tremolite rocks, marked by hornfelsic (in diopside varieties) and micronematoblastic structures (in tremolite varieties) and by massive textures.

A consideration of geologic relations of all these microgneisses and of their mineral and chemical composition suggests their primary tuffaceous sedimentary origin — in an alternation of fine-grained tuffs, polymictic varieties, and sedimentary rocks with addition of carbonaceous material. In contrast to other anticlinal folds of this anticlinorium, no porphyroids have been observed in this relatively depressed block.

Microgneisses carry stratified bodies of amphibolite. These are rocks with relict ophitic, diabase, and porphyritic textures, suitable as criteria in the differentiation of amphibolitic diabases and gabbro porphyries. Their conformable position, along with their participation in the folded structure and their absence in the overlying section, suggests their correlation with the tuffaceous sedimentary rocks.

Rocks surrounding the massif exhibit alternations expressed in their granitization (regional phenomenon) and in the albitization and microclinization in the outer contact zone (local phenomena).

Granitization is widely developed in rocks from the lower part of the Precambrian section in the Karsakpay region. It was expressed in the recrystallization of rocks, with the formation of coarser-grained textures accompanied by feldspathization, which has led to the formation of microgneiss, fine-grained granitoid gneiss, augen gneiss, and granodioritic gneiss — according to the composition of the original rocks. The granitized rocks are marked by a porphyroblastic texture, formed by coarse, amorphous new formations of plagioclase and K-feldspar in the matrix of a fine aggregate of minerals in paragneiss. The more granitized the rock, the lower its content of fine-grained aggregates of quartz and feldspar. Gradual transitions from fine-grained granitoid gneisses and augen gneiss to coarse-grained granitoid gneiss have been observed. The latter were described by Ye. L. Butakova as oligoclase-microcline to microcline-albite granites. The mineral composition of granitoid gneisses reflects the difference in the ratio of leucocratic to melanocratic minerals.

Granitoid gneisses are made up of plagioclase ranging from oligoclase-albite to albite No. 5-6, K-feldspar with perthite replacements, quartz, biotite or chlorite in various amounts, and muscovite. Accessory minerals are zircon, apatite, ore minerals, and sphene. A mixed-grained texture and a more or less gneissoid structure are typical; a cataclastic effect has been observed frequently.

Granites in the western part of the area are represented by massive equigranular rocks, close in composition to granitoid gneiss and connected with them by gradual transitions; at the same time, they underwent albitization expressed in the development of replacement perthite in K-feldspar. The granites are obviously palingenetic formations, because of their uniform hypidiomorphic texture.

The local manifestations of albitization in the outer contact zone of the massif have been superimposed on slightly granitized rocks represented by assorted microgneisses. Unlike the wide development of porphyroblasts in granitization, the metasomatic development of albite and microcline in microgneiss surrounding the syenite has led to the formation of a fine-grained aggregate of these minerals in the structural tissue of the rocks. For that reason, contact alterations are better seen under the microscope than in the field. However, a considerable albitization of microgneisses results in trachytoid textures formed by numerous small fibers (leists) of albite, up to 1.5 millimeters. Their development is very uneven and does not spread beyond a few tens of meters of the outer contact zone. A microscopic study of minerals in microgneiss reveals a great number of fiber-like (leist-like) bodies of fresh albite close in composition

and morphology to plagioclase in the albites.

The microclinization development is even more local. It is expressed in amorphous, lens-like, banded, and spotty leucocratic bodies, up to several centimeters in length, imbedded in the fine-grained matrix of microgneiss. Under the microscope, they display a wide development of slightly pelitic K-feldspar metablasts with an indistinct lattice. They are unevenly distributed throughout the rock, resulting in a glomeroblastic texture. Such bodies have no clean-cut outlines. Plagioclase in microgneiss was partially replaced by K-feldspar; consequently, it has microcline with a typical lattice developed on it.

Thus, a metasomatic development of K-feldspar was predominant in the granitization process, while albitization definitely prevailed in contact alterations near the syenite massif.

ALKALINE AND NEPHELINE SYENITES

The Karsakpay massif is made up of leucocratic and hastingsite syenite connected by gradual transition. Nepheline syenite and granosyenite occur in places in various parts of the massif, along with albitite in a thin body conformable with mesocratic gneiss. However, the detailed mapping of all these rock types is impossible, because they lie under the sodded course of a Sary-Say tributary. Only isolated small outcrops and very coarse sand occur within the massif.

Short dikes of leucocratic and hornblende syenite porphyries, 0.3 to 0.5 meters thick, have been observed in the massif. The first were traced by rock piles in its western part; the second were observed in an outcrop to the north, in a northeasterly trending dike and in scattered small chunks.

Leucocratic and hastingsite syenite are light-colored, pinkish to brownish and gray rocks with a variable amount of dark components — hastingsite and lepidomelane, both almost lacking in leucocratic alkaline syenite. The latter consist of microcline perthite and albite, with muscovite present in small amounts. Lepidomelane is present in small scales dispersed throughout the rock, and as inclusions in microcline perthite. When quartz is present, aggregates of lepidomelane are often associated with it. The amphibole type of hastingsite occurs in leucocratic syenite in small amounts, only, in very fine non-crystalline inclusions with evidence of corrosion by feldspar. Ilmenite is present to a considerable extent, among the accessory minerals, with small amounts of sphere, malacon (a variety of zircon), apatite, and orthite. When colored components are present, they show certain definite orientation and occur in aggregates.

The hastingsite varieties differ from leucocratic syenite in their sizable content of hastingsite and lepidomelane. The colored components occur in aggregates more or less evenly distributed in the matrix of feldspars. In such aggregates, the hastingsite grains are not uniformly oriented; in addition to the lack of any crystalline form, their sinuous outlines suggest a corrosion of amphibole by plagioclase. Individual, intensively corroded grains, exhibit some aggregates of uniformly oriented small inclusions separated by albite grains. Hastingsite displays a typically intense pleochroism, dark-gray along $\gamma = 1.740 \pm 2$; $\alpha = 1.721 \pm 2$. A chemical analysis of hastingsite by Ye. L. Butakova [3] indicates a high iron oxide content. Accessory zircon, orthite, and sphene are associated with hastingsite. The latter is partially replaced by lepidomelane with its intensive deep brown to light brown pleochroism.

In the leucocratic part of the rock, porphyroblastic perthite and chessboard albite form poikolitic growths with albite, of an irregular form and a polysynthetically twin structure. The albite aggregates display a tendency for the formation of coarser grains with a common orientation of the twin structure in poikolitic growths.

The relationship of nepheline and alkaline syenite under field conditions has not been ascertained because of their poor exposures. A microscopic study of these rocks reveals that their texture has a number of common features suggesting a common genetic series for both. Typical of nepheline syenite is the definite development of albite similar to that in other types of alkaline rocks, with a tendency for more idiomorphic inclusions, although they are devoid of crystallographic faces; such inclusions are lamellar (leist-like) to elongated prisms in shape. Sub-tabular porphyroblasts of chessboard albite, antiperthite, and nepheline, the latter partly replaced by muscovite and cancrinite, make up a considerable part of the rock. A colored mineral, apparently hastingsite, has been totally replaced by ore minerals and by a fine-scaled light green biotite. The bizarre outlines of these pseudomorphs suggest an intensive corrosion of the primary mineral by albite. In another variety of nepheline syenite, the colored hastingsite and lepidomelane have escaped the replacement but still display the corroded outlines. It is of interest that nepheline gravitates toward the aggregates of colored minerals and in places contains their relicts.

A special type of rock is represented by "muscovite" syenite where nepheline has been fully replaced by aggregates of muscovite and cancrinite, and its fine-grained texture is made up of lamellar (leist-like) albite, while the colored component is present in small amounts, only. Such varieties have a tracytoid texture discernible with the naked eye.

The extreme members of this series are albitites consisting almost fully of albite. They contain only small amounts of porphyroblasts of antiperthite, aggregates of muscovite, peculiar radial aggregates of later microcline, and relicts of hastingsite replaced by ore minerals and by small scales of light green biotite and muscovite. Lamellae of albite with a trachytoid texture and of varying size make a bizarre pattern about isolated porphyroblasts of antiperthite and pseudomorphic aggregates of muscovite.

Granosyenites are somewhat different in mineral composition and texture from the rest of the alkaline series; they have been observed in two separate outcrops. They are fully crystalline, medium-grained rocks composed of non-latticed K-feldspar with a perthitic texture, plagioclase, myrmekite, xenomorphs of quartz, and aggregates of colored minerals — hastingsite and biotite with orthite. A feature of these rocks is the presence of relict grains of sericitized plagioclase among the grains of K-feldspar and fresh No. 15 oligoclase. The grains of sericitized plagioclase in K-feldspar display a reaction fringe with specks and veinlets of calcite. Inclusions of biotite and isolated sericite aggregates are common in both K-feldspar and the fresh plagioclase. Amphibole, at its contact with K-feldspar, has been replaced by biotite. Thus, sericitic plagioclase and amphibole are relict minerals of a pre-existing rock, probably a mesocratic microgneiss.

Two types of vein syenite porphyry — leucocratic and hornblende — have been observed in the alkaline and nepheline syenite massif. Leucocratic syenite porphyry is a pinkish-gray, massive, fine-grained rock with small incrustations of feldspar, while the hornblende varieties carry a large amount of acicular hornblende, noticeably oriented parallel to the dike contacts, and responsible for the green-gray color of the rock. The porphyroid texture of both types stands out under the microscope — expressed by sharply zoned plagioclase and a few isometric inclusions of hornblende, in the leucocratic type; and by the abundance of acicular hornblende crystals, up to 3 millimeters, in the hornblende type.

Plagioclase incrustations in the leucocratic syenite porphyry are interesting because of their abnormal zone of intermediate plagioclase marked by a refractive index lower than that for the plagioclase on either side of it. Plagioclase in the inclusions is partly replaced by sericite. The granular rock matrix consists of plagioclase, K-feldspar, and small dispersed scales of biotite or amphibole. The outer fringe of albite about the intermediate-size plagioclase grains is serrate, with the indentations filled by small grains of the groundmass.

The hornblende syenite porphyries carry acicular bodies of hornblende, progressively decreasing in size, with typical crystallographic cross sections, and also aggregates of isometric feldspar grains. Biotite and sphene in association with amphibole are present to a small extent, along with epidote. Accessory apatite is present. The rock contains K-feldspar, sericitic plagioclase with a refractive index close to that of Canada balsam, and fresh albite which is developed in bands and spots. When present in small amounts, amphibole is not subject to corrosion. However, other varieties of these vein rocks show considerable alterations in composition and texture, with increasing importance of fresh albite.

Albitization is widely developed in all rocks of the massif. It has controlled a continuous series of changes in the composition and texture of alkaline rocks — from leucocratic syenite to albitite. Especially clear are the successive stages of albitization in the group of leucocratic alkaline syenites. The more or less even- to coarse-grained varieties, made up to a considerable extent of microcline-perthite, change to varieties with very unevenly grained poikiloblastic and corrosion textures, with a wide development of albite (Figure 3).

In leucocratic alkaline syenite (Figure 3-a), microcline-perthite represents the decay perthites — thin, regularly-oriented growth of albite No. 2-4 (2V-84 to 88) in K-feldspar. Located among coarse tabular bodies of microperthite are chains of the same albite but without the twin structure. However, such rocks occur very rarely in the massif. Even a single thin section may show an initial stage of the change from distinctive fine perthite growths to coarse pellicular to vein type replacement perthites. The latter are common in leucocratic alkaline syenite. They show double fringes of albite between two adjacent bodies of microcline-perthite, giving the impression of a mutual intergrowth of alternately oriented albite growths (Figures 3-b, 4). These grains of albite exhibit a twin structure in places. An increase in albite results in the formation of double chains of grains, each showing the same extinction, and separated by a chain of albite grains with a different extinction.

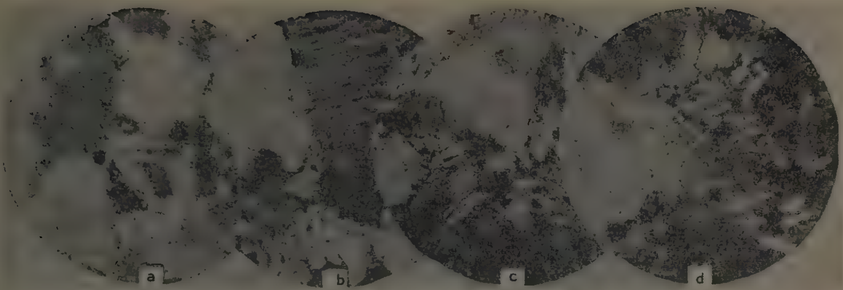

FIGURE 3. Development of albitization and microstructure in alkaline syenite. Magnification 27X, with analyser.

This phenomenon was noted by Ye. L. Butakova [3] in her description of the Karsakpay syenite; by V.S. Sobolev [8], for microcline-perthite granitoids of the Korosten' pluton; and by L.F. Aynberg [1] for syenite from the Azov alkaline massif. V.S. Sobolev ascribes their origin to the unmixing of solid solutions. A detailed description of the double fringes and double bands was given by S.F. Vasil'chenko [4] for plagioclase from hornblende gneissoid rocks of a contact-metamorphic origin. He associates this phenomenon with diablastic formations. In the syenite we have studied, the double fringes and bands in microcline-perthites occur chiefly in rocks with a wide development of replacement perthite, and obviously are also of a diablastic origin.

Besides the double fringes, a fine-grained aggregate of albite occurs among the tabular inclusions of microcline-perthite, which it corrodes (Figure 3-b, c) making them assume irregular sinuous outlines. The development of replacement perthite brings about the formation of spotty perthite with fine polysynthetic twin structure, and even the formation of chessboard albite with small relicts of microcline. This microcline is pelitic and shows a spotty extinction. In the presence of a considerable development of albite, it usually has a twin structure and a tendency to idiomorphism with the formation of elongated prismatic bodies without terminal faces. The albite grains have a somewhat higher refractive index than the perthitic albite growths, and approach albite-oligoclase in composition. An overgrown aggregate of such grains produces complex microcline growths with an unevenly grained poikiloblastic texture (Figure 3-d).

Thus, leucocratic syenite displays microstructural varieties, from equigranular coarse microcline-perthite, to a definitely mixed-grained type with clean-cut corrosion and poikiloblastic growths. Simultaneously, albite No. 2 to 4 change to albite 7 to 8, with the proportion of plagioclase increasing, compared with K-feldspar. Structurally, such varieties are close to hastingsite syenite.

Typical of the latter is the development of poikiloblastic relations between the porphyritic inclusions of antiperthite and chessboard albite, and the finer-grained aggregate of polysynthetically twinned albite, approaching albite-orthoclase in composition. Seen in segments with these aggregates of grains is the common orientation of couble bands in separate grains. Corrosion forms are very characteristic of the hastingsite inclusions; they are represented by aggregates of variously oriented drop-shaped to rounded or polygonal grains, utterly devoid of crystallographic faces. Hastingsite has been more or less replaced by lepidomelane.

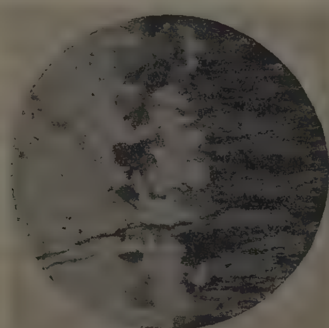

FIGURE 4. Double chains of albite in leucocratic alkaline syenite.

Magnification 52.5X, with analyser.

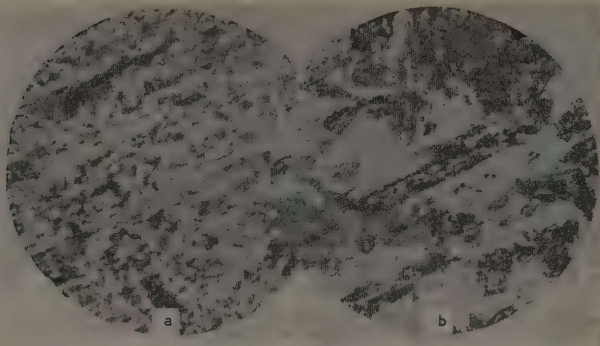

FIGURE 5. Non-albitized (a) and albitized (b) hornblende syenite porphyry.

Magnification 52.5X, with analyser.

Nepheline syenite shows the same albitization as hastingsite syenite, with a corrosion of nepheline by albite. Albitization is especially complete in albitites which are almost pure albite.

It is of interest that intensive albitization is present locally in vein rocks of the syenite massif, where it is expressed in an increase in fresh albite as compared with sericitic albite-oligoclase in slightly altered varieties; in the appearance of microcline-perthite with veinlets and spots of replacement perthite, leading to the formation of chessboard albites; and in the development of poikiloblastic structures similar to those noted in syenite, and to a smaller extent of double fringes of albite. Very typical is the active relationship of albite and acicular amphibole, which goes as far as the formation of hornblende "skeletons" (Figure 5).

Standing out against the background of intensive albitization in rocks of the massif and of the vein series, some varieties of leucocratic syenite exhibit vague vein-line segregations of a younger and purer microcline with a clean-cut typical lattice and without any perthite growth. The replacement of amphibole by biotite and the development of biotite suggest the participation of potassium in the metamorphic process.

The youngest formations are fine veins of albite locally developed in shattered rocks which display a small displacement of double bands in relatively large albite segregations; in addition, the bands are cut by small transverse albite veins with a lower refractive index (Figure 6).

A comparison of all rocks of these types shows the increasing importance of albite in the series of leucocratic alkaline, hastingsite, and nepheline syenites and albitites; a change in the morphology of the albite grains, from fine and isometric to elongated lamellar (leist-like); the appearance of a more or less definite twin structure; an increase in the size of inclusions, from the hundredths to 2 millimeters; an alteration of microcline-perthite, from decomposition perthite to antiperthite and chessboard albite, through assorted replacement perthites; and a wide development of còrrosion, poikiloblastic textures.

In addition to this well-expressed metasomatic development of albite, there is indirect evidence of metasomatic formation of nepheline, through nephelinization of pyroxene.

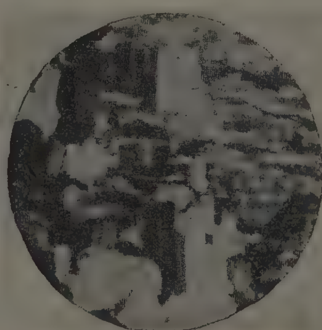

FIGURE 6. Development of younger albite in connection with breaking up of syenites.

Magnification 27X, with analyser.

This is suggested by the latter's association with aggregates of dark minerals and by their relicts in nepheline. The appearance of calcite and coarse grains of apatite in such varieties points to a local manifestation of calcium metasomatism, in connection with the replacement of pyroxene by nepheline. In the Karsakpay alkaline and nepheline syenite massif, these phenomena are apparently related to a metasomatic alteration of pyroxene in gabbro-amphibolite xenoliths. Such phenomena are not common; they are similar to metasomatic processes of pyroxene nephelinization in massifs of ultrabasic to alkaline rocks described by L. S. Borodin [2].

THE ORIGIN AND AGE OF THE KARSAKPAY ALKALINE MASSIF

The development of widespread albitization in alkaline rocks of the Karsakpay area, as compared with the granites, is obvious in chemical analyses. The alkaline content increases from leucocratic alkaline to nepheline syenite. Data of chemical analyses as well as the variation diagrams from the four analyses of alkaline and nepheline syenites on hand (Figure 7) show that the numerical characteristic a increases sharply; as seen from the ratio of weight percent, for Na_2O and K_2O, this increase is connected with an appreciable increase in N_2O compared with K_2O. The K_2O content varies in the 5. 7 to 4. 2% (weight) range, and is on the whole close to that in granite. With a noticeable décrease in the importance of microcline in microcline-perthite of alkaline rocks, such a sizable K_2O content is due apparently to the appearance of younger and fresher microcline, as a result of redeposition.

The alkaline and nepheline syenites under study have a relatively low value of numerical characteristic c, being poor in CaO and supersaturated by alumina. The numerical characteristic b, comparatively high in alkaline and hastingsite syenites, increases mostly with an increase in iron oxides in ilmenite (FeO), and of amphibole in hastingsite syenites. In a variation diagram, numerical characteristic s reflects a decrease in the silica content in the alkaline rock series.

A comparison of chemical analyses of the Karsakpay granite and syenite, reveals a fairly regular increase in parameter a and a decrease in parameter s, in the granite — alkaline syenite — nepheline syenite series. The presence of granosyenite within the syenite massif, and the proximity of leucocratic alkaline syenite to the granite series rocks, also suggest a genetic relationship of syenite and granite. As already noted, the granite, too, is affected by albitization as expressed in the formation of microcline perthites; however, its effect here is less intensive.

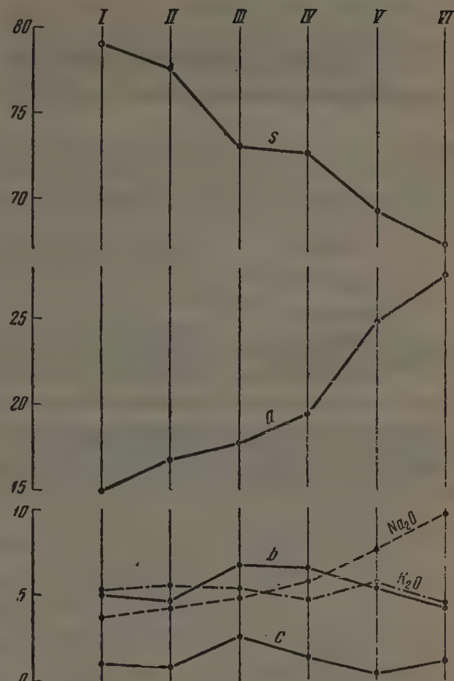

FIGURE 7. Variation diagrams for chemical composition of granites and syenites.

Numerical data obtained by the A.N. Zavaritskiy method.

Metasomatic phenomena have affected rocks of the massif to such an extent that the composition of the primary igneous rocks can only be surmised. However, the presence of xenoliths of gabbro-amphibolites and hastingsite; of biotite in aggregations; of quartz in association with these minerals, in leucocratic and even hastingsite syenite; of relict plagioclase in granosyenite; and of irregular zoned incrustations of plagioclase in vein syenite porphyry — all suggest a contamination of the magma during formation of the syenite massif. The presence of quartz-diopside and quartz-tremolite rocks in the section, along with gabbro-amphibolites, may have promoted the enrichment of the granite magma in such components as CaO, MgO, and FeO (reflected in the relative rise of parameters c and b in variation diagrams for leucocratic and hastingsite syenites). This evidence of contamination of the magma

Table 1

Chemical composition of granite and syenite from the Karsakpay Area (in weight %)

Sample No.	Rock	Components											
		SiO	TiO_2	Al_2O_3	Fe_2O_3	FeO	MnO	MgO	CaO	Na_2O	K_2O	$H_2O(\pm)$	total
I	Granite, say Baykonur	73,70	0,35	13,71	2,27	0,22	0,06	0,15	0,38	3,68	5,26	0,32	100,10
II	Granite, say Aktas	70,60	0,13	16,53	0,84	0,75	trace	0,04	0,64	4,19	5,59	0,48	99,79
III	Syenite	65,56	0,33	17,27	1,78	3,08	0,12	0,10	1,16	4,76	5,34	0,40	99,90
IV	Hastingsite syenite	64,14	0,32	18,60	2,44	1,60	0,10	0,12	1,15	5,70	4,84	0,84	99,85
V	Hastingsite syenite	61,44	0,20	19,30	1,61	2.92	0,11	0,13	1,24	7,61	5,66	0,54	100,76
VI	Nepheline syenite	58,90	0,04	24,13	0,81	0,29	0,10	0,20	0,87	9,70	4,24	1,04	100,32

Analyses performed: I -- in chemical laboratory of G.I.N. (Geol. Institute) Ac. Sc. U.S.S.R.; Ye. A. Shilova, analyst; II and V -- in the chemical laboratory of Inst. of Geology of Ore Deposits, Petrography, Mineralogy, and Geochemistry (I.G.E.M.); M.N. Venprintseva, analyst; III -- chemical laboratory G.I.N.; A.N. Zarubitskaya, analyst; IV and VI -- after Ye. L. Butakova [3].

Table 2

Numerical characteristics by the A.N. Zavaritskiy Method

	I	II	III	IV	V	VI
a	15	16,9	17,7	19,4	24,9	27,5
c	1	0,8	2,5	1,4	0,3	1.1
b	5	4,7	6,7	6.5	5.5	4.1
s	79	77,6	73.1	72.7	69.3	67,3

NOTE: Comma represents decimal point.

is hardly compatible with a deep nature of the latter.

The appearance of albitization in alkaline rocks is obviously related to a structural position different from that of the Karsakpay granites. The latter have been traced discontinuously for over 100 kilometers along the western rim of the Maytyubin anticlinorium, with a maximum width of the outcrops of 6 kilometers. They appear to be associated with a pervious zone of ancient meridional faults in that part of the anticlinorium. The subsequent resumption of tectonic movements in that zone brought about post-intrusive submeridional dislocations, as a result of which it became associated with troughs of Middle Paleozoic rocks. The syenites, on the other hand, are localized in the folded structure of the anticlinorium, whose origin, to judge from facies differences in the middle formation of the lower Proterozoic, also by the angular and azimuthal unconformities and by the presence of conglomerate higher up in the section, was dated as earlier than the deposition of the upper Proterozoic sequence. The formation of this massif in an apparently slightly pervious anticlinal structure favored the development of intensive albitization. Later faulting, trending northwest to near-latitudinal, has led to the formation of a block including the fold within the syenite massif.

The occurrence of alkaline and nepheline syenites is unusual in the Maytyubin anticlinorium; it is due probably to a combination of favorable structural and lithologic features of rocks reacting with a granitoid magma.

Structural differences in the geologic position of the Karsakpay granite and syenite suggest a different nature of tectonic movements in the process of intrusion. On the other hand, their genetic relation — as noted before — may confirm A.N. Zavaritskiy's idea on the effect of tectonic movements in the earth's crust on magmatic processes: "It is not the

primeval differences in a magma that account for the difference in complex petrographic compounds originating from it; different compounds may originate from the same magma, under different conditions of its evolution" ([7], p. 26).

The Karsakpay alkaline and nepheline syenites may be correlated with the Il'men miaskites and alkaline syenites [6], in petrographic composition, their association with paragneisses and granitoid gneisses, their genetic relation to the granite magma, and their geologic position in the folded structure. The wide development of metasomatic processes, noted by A. V. Vlasenko [5] for an alkaline complex in the south Il'men Mountains, with a similar replacement of Na-plagioclase by a fine-grained aggregate and by a younger fine-grained, latticed feldspar, is also very much like the phenomena which we have noted in the Karsakpay massif.

Unfortunately, there are no direct data on the age determination of the Karsakpay massif as distinct from other Precambrian rocks and ancient granitoid gneisses.

Ye. L. Butakova [3], in noting the fresh appearance of mineral components and the absence of any evidence of any strong cataclastic phenomena, correlates the intrusion time with the last stage of an intensive folding, believing that such a folding in this part of Kazakhstan was Variscan. She assigns the gneissoid granites to Precambrian formations, tentatively assigning a Caledonian age to the basic intrusions (gabbro-amphibolites). According to present concepts, the western part of central Kazakhstan is a Caledonian fold province, and no younger intrusions have been observed south of Ulutau. As noted before, the gabbro-amphibolites are syngenetic with the lower, Precambrian, sequence. Cataclastic phenomena in syenites have been noted by B. S. Dubova and the author.

K. A. Rachkovskaya notes the gradual transition from granites to syenites, and a considerable fragmentation of both. She assigns the syenites to formations close to granite, in age, and originating from the same magma. We have not observed any gradual transition from granitoid gneisses to syenites; however, both rocks have certain similar features, such as the development of microcline perthite with replacement perthite, up to the formation of chessboard albites; also the appearance of fresh microcline. At the same time, the granitoid gneisses clearly display the features of genesis as a result of granitization. As pointed out before, a number of features suggests a genetic relation between alkaline and nepheline syenites, and those granites which are distributed in the same area, west of the syenite. The age position of the granites is definitely established by the presence of their pebbles, along with those of granitic Precambrian rocks, in upper Proterozoic (upper course of, perhaps, the Tumurza) and Cambrian (Bassaltuyat Mountain) rocks.

It may be assumed, then, that the Karsakpay alkaline and nepheline syenites are Precambrian formations representing a special type of alkaline intrusions in a folded province.

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Geological Institute,
Academy of Sciences U. S. S. R.,
Moscow.

Received May 15, 1958.

MINOR INTRUSIONS IN DZHUGA MOUNTAIN AND IN THE BASIN OF KISHA AND BEZYMYANNAYA RIVERS[1]

by

V. M. Ayanov

GEOLOGIC ENVIRONMENT AND AGE OF THE INTRUSIONS

Igneous rocks, known as "minor intru-sions," are distributed in the western part of the northwestern Caucasus' intermediate zone, in Dzhuga Mountain and along Kish and Bezymyannaya Rivers (Fig. 1). They are described in papers of Ye. N. D'yakonova-Savel'yeva [3], A. G. Kobilev [6], V. A. Zatokovenko [4], G. A. Afanas'yev [1], and A. A. Kadenskiy [5].

Field observations have revealed the presence of two different intrusive complexes in this area. The older of the two is represented by granitoid rocks distributed mostly east of the Dzhuga where they make up the Chelepsy Range. Among the Chelepsy rocks, A. G. Kobilev [6] has identified a large granite intrusion with chloritic biotite, while G. D. Afanas'yev has described [1] granodiorite and alaskite — typical fully-crystalline deep-seated formations considerably altered by secondary processes.

The Chelepsy granitoid intrusions are pre-Lower Permian and obviously pre-Middle Carboniferous, because Lower Permian deposits rest directly on them, while Middle Carboniferous deposits, resting on lower Paleozoic contact metamorphic schists which enclose the intrusions, have a normal sedimentary aspect.

These granitoids enclose rocks of a younger intrusive complex, here described as the minor intrusions, and developed chiefly in Dzhuga Mountain. Identifiable here are diorite at its summit, granodiorite and quartz diorite in the northeastern spurs, and leucocratic granite and granitic aplite on the south slopes and cutting the diorite and granodiorite (Fig. 2).

Diorite at the top of the Dzhuga form a small (less than a square kilometer) stock piercing lower Paleozoic schist and the older Chelepsy Mountain granitoids.

A minor intrusion in the northeastern spurs also is stock-like, elongated meridionally. It contains, besides amphibole-bearing granodiorite, quartz diorite which is exposed in the periphery. There are gradual transitions from granodiorite to quartz diorite. The most acid members of this association — leucocratic granite and granitic aplite — occur in dikes and veins. They cut the diorite, quartz diorite, and granodiorite at the top of the mountain and in its northeastern spurs. All intrusive rocks, in their turn, are cut by barite and quartz veins with evidence of polymetal mineralization. The trend of hydrothermal veins coincides on the whole with the northeastern trend of leucocratic granite dikes.

An idea of the upper age boundary of the Dzhuga area minor intrusions can be obtained from the following facts. Their hydrothermal derivatives — barite veins — are traceable in Lower Permian redbeds (P_1^1), transgressive on ancient granitoids of the Chelepsy Range, east of the Dzhuga. These veins are absent in the overlying Lower Permian gray beds (P_1^2), separated from the redbeds by an angular unconformity. Consequently, the formation of minor intrusions, including their hydrothermal derivatives, had been completed prior to the second half of the Early Permian.

A small outcrop of similar granodiorite and quartz diorite is located along Kisha River, north of Kholodnaya River (left tributary of the Kisha). Here, the minor intrusion is confined by sandstone and conglomerate — Upper Carboniferous, according to V. N. Robinson [7]. In an exposure in the right bank of the Kisha, apophyses of a granodiorite intrusion penetrate Upper Carboniferous conglomerate, and carry xenoliths of the wall rock. The contact line is broken by numerous normal faults which have dropped sedimentary blocks deep into the intrusive body (Fig. 3).

Minor intrusions also crop out in a small

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[1]Malyye intruzii gory Dzhuga i basseyna rek Kishi i Bezymyannoy.

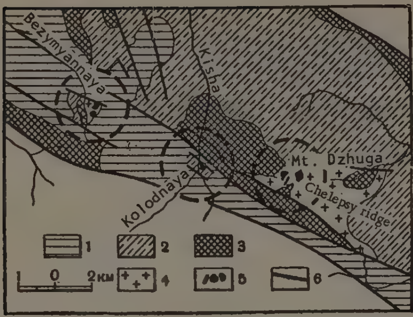

FIGURE 1. Areas of minor intrusions in the intermediate zone of northern Caucasus.

1 -- Mesozoic; 2 -- Upper Paleozoic; 3 -- undifferentiated Lower and Middle Paleozoic; 4 -- granitoids of the Chelepsy type; 5 -- minor intrusions; 6 -- faults.

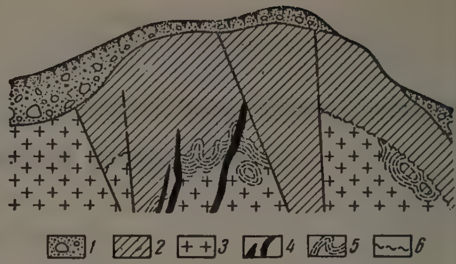

FIGURE 3. Minor intrusion exposed on the right bank of Kisha River.

1 -- Diluvium; 2 -- Upper Carboniferous conglomerate, sandstone, and less common shale; 4 -- hydrothermal formations; 5 -- flow structure; 6 -- shattered zone.

crystalline massif of Bezymyannaya River. As early as 1938, A. G. Kobilev [6] pointed out a different age for the intrusives of that massif. The older of these are represented by granite of the Chelepsy type; the younger, by quartz diorite occurring in a small body in the northern part of the massif. Identifiable besides the quartz diorite here are diorites similar to those from Dzhuga Mountain, and vein granite close in composition to the Dzhuga type leucocratic granite. The diorite forms a small stock isolated from the quartz diorite outcrops. Vein and leucocratic granites cut all other intrusives of Bezymyannaya Mountain.

On the basis of the foregoing, the age interval of these minor intrusions lies between the end of the Late Carboniferous and the second half of the Early Permian. The diorite, being the earliest intrusive phase, appears to have been formed also at the close of the Carboniferous.

On the whole, the intermediate zone intrusives are upper Paleozoic.

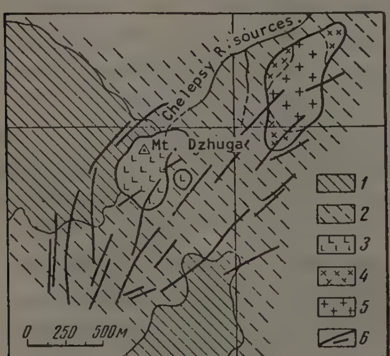

FIGURE 2. Outcrops of minor intrusions in the Dzhuga Mountain area.

1 -- Lower Paleozoic metamorphics; 2 -- ancient granitoids of the Chelepsy type; 3 -- diorite of minor intrusions; 4 -- quartz diorite of minor intrusions; 5 -- granodiorite of minor intrusions; 6 -- leucocratic granite and granitic aplite of minor intrusions.

PETROGRAPHY OF PRINCIPAL TYPES OF INTRUSIVE ROCKS

The Dzhuga Mountain diorites, externally, are fine- to medium-grained dark-green rocks, porphyritic in the peripheral parts, with porphyroblasts of feldspar.

The principal rock-forming minerals are plagioclase and amphibole, with accessory K-Na-feldspar and biotite, and secondary chlorite, epidote-zoisite, and sericite.

Plagioclase — andesine No. 40 (with the albite twinning law dominant) — forms idiomorphic, slightly sericitic crystals. The isolated crystals of K-Na-feldspar show a poorly expressed microcline lattice and an inconsistent optic angle (2V = -60 to -75°).

Hornblende is represented by relatively idiomorphic prismatic crystals. The γ pleochroism is brownish-green with a blue tine; the β is green; the α is light yellow-green.

Angle 2V = -60 to -64°; $c\gamma$ = 18°. Refractive indices: γ = 1.680; α = 1.663 ($\gamma - \alpha$ = = 0.017). Biotite usually occurs in fringes about some hornblende grains. It is greenish-brown along γ.

Accessory minerals are apatite, sphene, zircon, ilmenite, and magnetite. Chlorite is developed to the same extent on amphibole and biotite, the replacement being slight in both minerals.

The Bezymyannaya Mountain diorite differs little from that of Dzhuga Mountain (Table 1) and is only slightly more affected by secondary processes.

The chemistry of these diorites (Table 2, specimen 1574) is marked by the deviation in composition from the normal — the intermediate, according to R. Daly — diorites in the direction of more alkaline and melanocratic rocks.

Quartz diorites of Dzhuga Mountain and Bezymyannaya River are green-gray, mixed-grained, commonly porphyritic rocks. They consist of quartz, plagioclase, K-Na-feldspar, amphibole, and biotite, with accessory apatite, sphene, zircon, and magnetite, and secondary chlorite, epidote-zoisite, and sericite.

The quartz is coarsely crystalline, with grains up to 3 millimeters, allotriomorphic, with a cloudy extinction.

Plagioclase is andesine No. 35-38 (with the albite and pericline twinning predominant). Idiomorphic crystals, sericitic along the cleavage planes, are in places zoned, with the outer and more acid fringe represented by No. 25-28 oligoclase.

The K-Na-feldspar occurs as latticed and non-latticed (angle $\beta \perp (001)$ = 12°) microcline with a variable optic angle, 2V = 68 to 82°.

The common hornblende commonly forms simple twins. Pleochroism along γ is deep brown-green; along α, light yellow-green. The optic angle (2V) = -64 to -68°; $c\gamma$ = 18°. Birefringence, $\gamma - \alpha$ = 0.020 – 0.022.

Biotite forms individual tabular crystals, up to 2 millimeters along the long axis, colored deep greenish-brown along γ.

Chlorite is developed on amphibole and to a smaller extent on biotite.

There are isolated instances of leucoxene fully replacing ilmenite crystals.

The Kisha River quartz diorites are similar to those described above, differing from them in a higher microcline content, up to 4.2%, in the more intensive chloritization of colored minerals, and in stronger chloritization of plagioclase.

The quartz diorites differ also in their higher alumina content, and by subordinate magnesium in magnesial-ferruginous components.

Table 1

Mineral Composition of Rocks

Rocks*	Quartz	Plagioclase	K-Na-feldspar	Biotite	Muscovite	Amphibole	Apatite	Sphene	Ore-bearing	Secondary (epidote-zoisite, chlorite)
Diorite, Dzhuga Mt.	0,1	52,3	2,1	2,3	—	40,4	0,3	1,0	1,4	0,1
" , Bezymyannaya R.	0,6	50,8	1,6	3,0	—	38,2	0,5	0,8	1,5	3,0
Quartz diorite, Dzhuga Mt.	14,0	53,5	2,0	11,5	—	16,7	0,6	0,2	1,0	0,5
" Bezymyannaya R.	12,5	54,0	2,5	10,0	—	15,0	0,5	0,1	0,5	4,9
" Kisha R.	15,0	52,0	4,2	8,6	—	12,0	0,5	0,5	0,8	6,4
Granodiorite, Dzhuga Mt.	23,0	50,2	6,7	8,3	—	10,8	0,3	—	0,4	0,3
" , Kisha R.	21,0	50,0	6,5	9,0	—	11,5	0,5	—	0,2	1,3
Leucocratic granite, Dzhuga Mt.	31,2	36,5	28,1	2,8	0,9	—	—	—	0,3	0,2
" Bezymyannaya R.	30,5	36,8	29,0	2,5	0,6	—	—	—	0,2	0,4
Granite-aplite, Dzhuga Mt.	30,3	36,2	31,8	0,8	0,9	—	—	—	—	—

*Average from 10 to 15 computations for each rock.

NOTE: Comma represents decimal point.

The Dzhuga Mountain granodiorite externally is similar to quartz diorite, differing from quartz diorite in a lighter greenish-gray color and in comparatively uniform medium-size grains. It consists of quartz, plagioclase, amphibole, biotite, accessory K-Na-feldspar, and secondary chlorite, the epidote-zoisite group minerals, and sericite.

Quartz forms xenomorphic crystals with a cloudy extinction; it also occurs in poikilitic growths in amphibole and plagioclase.

The plagioclase is oligoclase with an average of 25% of the anorthite molecule. Rocks from the transitional granodiorite – quartz diorite zone contain a more basic plagioclase (An, up to 30%). Albite is developed on the periphery of feldspar crystals.

The K-Na-feldspar is microcline (2V = =-62 to 65°, and 70 to 72° in rocks from the central parts of intrusive bodies).

The amphibole is common hornblende, with pleochroism (biotite absorption) drab green, at times with a faint bluish tint, along γ; green along β; and greenish-yellow along α; $c\gamma = 17-20°$; $2V = -62-64°$; $\gamma - \alpha = 0.018 - 0.022$.

Biotite pleochroizes up to deep brown with a greenish tint, along γ, commonly displaying definite biaxiality. Its elongated tabular crystals often carry inclusions of radioactive zircon surrounded by pleochroic areas with a radius of 0.012 millimeters.

Accessory minerals are apatite, sphene, zircon, and magnetite.

Chlorite is formed at the expense of amphibole and biotite, with the hornblende usually replaced to a greater extent than biotite.

Under the microscope, the Kisha River granodiorite shows a considerable degree of alteration: its plagioclase is strongly sericitic; amphibole is in places fully replaced by chlorite, fine-grained epidote, and calcite; biotite forms aggregates among the other minerals; and submicroscopic scales of chlorite and streamlined secondary quartz are developed in fractures.

Leucocratic granite, externally, is a mixed-grained, mostly fine-grained, light-gray rock.

The principal rock-forming minerals are quartz, plagioclase, and K-Na-feldspar, with accessory biotite, muscovite, calcite, chlorite and epidote-zoisite.

Quartz is commonly intergrown with feldspars.

The plagioclase is represented by albite-oligoclase or else an acid oligoclase (An, 10 to 20%). Its idiomorphic crystals are slightly sericitic.

The K-feldspar is microcline [angle $\beta \perp (001) = 10$ to $14°$; $2V = -56$ to $52°$].

Biotite forms brown-green plates, 0.2 to 0.3 centimeters, in their longest dimension. Muscovite occurs in a parallel growth with biotite. Calcite has been observed in aggregates of fine grains among the other minerals.

Accessory minerals are apatite, zircon, occasional garnet, and magnetite.

Chlorite, developed either on biotite or independently, appears in sizable amounts in the Bezymyannaya leucocratic granite.

A feature of the chemical composition of the Dzhuga Mountain leucocratic granite (specimen 1623-a) is its higher alkalinity and lime content.

Granitic aplite has been observed only in the Dzhuga area. Macroscopically, this is an extremely fine-grained rock, pinkish to almost white, with an aplitic texture.

Principal rock-forming minerals are quartz, plagioclase, and K-Na-feldspar, accounting together for 99% of the total. Subordinate minerals are muscovite and biotite; accessories – apatite, zircon (rare), garnet, and magnetite.

The quartz shows in places a tendency for idiomorphism.

Plagioclase is represented mostly by albite-oligoclase (An, 10 to 15%), with finely sericitic crystals.

The K-Na-feldspar is a microcline, either latticed or non-latticed [angle $\beta \perp (001) = 10$ to $14°$], with a small optic angle ($2V = 60 - 65°$).

Biotite appears in small scales in the peripheral parts of aplite vein bodies. Muscovite, on the other hand, is evenly distributed throughout the rock. Accessory minerals are of little importance in granitic aplite rocks.

All these rocks are the main igneous rocks proper, present as minor intrusions. They are the most common and mineralogically and structurally most consistent, not only within a single locality but in areas distant from each other, as well.

Table 1 shows the results of a statistical computation of the quantitative mineral composition for main types of intrusive rocks

from different areas of the intermediate zone.

As seen from the petrographic description, diorite, quartz diorite, and granodiorite are mineralogically similar, as witness the identical optical properties of their hornblende and the consistency in the composition of their accessories. Both leucocratic granite and granitic aplite have the same high alkaline feldspar content, and are low in colored and accessory minerals.

A comparison of the chemical compositions of these rocks (Table 2) shows two sharply differentiated series of intrusives: diorite-granodiorite and granite.

Diorites-granodiorites. A common feature of the chemism of diorites, quartz-diorites, and granodiorites is their comparatively low absolute content of silica as against a high content of alumina, ferro-magnesial, and calcareous components. The chemical composition of diorites and granodiorites is

Table 2

Chemical Composition of Rocks from Small Intrusions, and Its Basic Numerical Characteristics (After A.N. Zavaritskiy)

Components	Dzhuga Mt. diorite, Spec. 1574	Bezymyannaya Mt. quartz-diorite, Spec. 239-a	Dzhuga Mt. Granodiorite Spec. 1623-b	Dzhuga Mt. leucocratic granite, Spec. 1623-a	Dzhuga Mt. granite-aplite
SiO_2	51,94	56,38	64,36	72,12	71,37
TiO_2	0,94	0,63	0,12	0,04	0,04
Al_2O_3	17,11	18,29	16,06	12,13	14,34
Fe_2O_3	0,96	0,72	0,76	0,09	1,29
FeO	5,03	4,88	3,30	1,36	0,32
MnO	0,05	0,06	0,05	0,02	0,02
MgO	5,03	2,38	1,90	0,11	0,23
CaO	6,90	5,43	2,67	3,40	2,08
Na_2O	3,96	3,30	4,30	3,70	3,41
K_2O	1,82	1,90	2,50	4,21	4,38
P_2O_5	0,14	0,23	0,12	0,01	0,02
SO_3	0,44	0,14	0,14	0,05	0,19
Loss on ignition	5,08	4,93	2,46	2,24	1,99
Total	99,41	99,32	99,26	100,30	99,68

Table 2, Continued

Numerical characteristics	Spec. 1574	Spec. 239-a	Spec. 1623-b	Spec. 1623-a	Spec. 1623
a	12,0	10,8	13,2	13,9	14,0
c	6,2	7,2	3,3	1,0	2,6
b	17,8	11,5	9,0	4,6	11,9
s	63,9	70,4	74,4	80,4	81,4
a'	—	11,7	20,0	—	—
f'	33,4	50,0	43,8	28,5	78,5
m'	51,4	38,3	36,1	4,3	21,4
c'	15,1	—	—	66,6	—
n	77,0	72,6	73,0	56,7	53,9
φ	5,0	5,2	7,7	1,4	14,2
t	1,2	0,8	—	—	—
Q	−2,5	+12,1	+19,2	+32,1	+32,3
a/c	1,9	1,4	4,0	13,9	5,4

NOTE: Comma represents decimal point.

marked by a higher alkali content.

Thus diorites, quartz diorites, and granodiorites may be regarded as related magmatic formations whose development proceeded in two stages — a diorite phase and a phase of granodiorites and quartz-diorites (Figure 4).

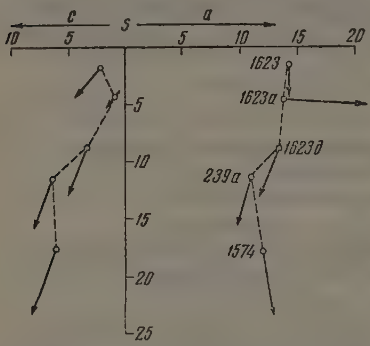

FIGURE 4. Composition diagram for minor intrusions in the intermediate zone, northwestern Caucasus.

Granites. Leucocratic granites and granite-aplites stand out in their almost identically higher content of silica and alkalis. They represent probably the residual melt of a deep-seated granitoid magma. It is not impossible that this residual melt, in further differentiation, gave rise to a comparatively independent sub-phase of aplite granites.

METAL CONTENT OF MINOR INTRUSIONS

The haloes of the Dzhuga Mountain granite intrusions contain many shows of barite, either with or without a polymetal mineralization. They are hydrothermal veins with barite, quartz, and calcite for principal minerals, and with galena and chalcopyrite for the accessories. The veins are associated with the same fault system as the leucocratic granite dikes. Because of this structural relationship, and also because the minor intrusions granitoids are the only igneous formations almost contemporaneous with the hydrothermal phenomena, we regard the latter as derivatives of the former. This is confirmed by a spectrographic study of the minor intrusions.

Table 3 lists the data on the content of some elements in main types of intrusive rocks.

Copper occurs in granodiorites and leucocratic granites, in an amount higher than that for an ore-free mineralization (according to I. I Ginzburg, [2]). It does not form independent minerals, being apparently dispersed throughout the rock. In mineral-forming amounts, it is concentrated in hydrothermal formations.

Table 3

The Content of Some Elements in Rocks of Minor Intrusions, and the Principal Ore-Making Minerals in Hydrothermal Formations

Elements*	Rocks of minor intrusions				Ore-free mineralization (2)	Ore minerals in hydrothermal derivatives of minor intrusions
	diorite	granodiorite	granite	granite-aplite		
Ti	0,650	0,260	0,020	0,015	—	
Co	0,003	—	—	—	—	
V	0,007	0,006	0,004	0,002	To 0,030	
Cu	0,006	0,009	0,010	0,003	0,006	Chalcopyrite
Zn	<0,01	<0,01	—	—	0,010	
Pb	0,005	0,010	0,009	0,007	0,006	Galena
Zr	0,012	0,020	0,015	0,005	—	
Be	0,003	0,005	0,009	0,010	—	
Ba	0,080	0,130	0,200	0,340	0,060	Barite
Sr	0,060	0,040	0,034	0,030	0,060	

*Averages of 25 spectrum analyses of diorites; of 25 for granodiorites; 18 for granites (leucocratic); 12 for granite-aplites.

NOTE: Comma represents decimal point.

The lead content in the intrusive rocks, with the exception of diorites, is above that for an ore-free mineralization. No independent lead ore minerals have been found. This element probably forms an isomorphic addition in feldspars and micas, and enters the hydrothermal phase as galena.

The barium content, in all rocks, is higher than that for an ore-free mineralization; it increases from intermediate to acid rocks, attaining its maximum in leucocratic granites and granite-aplites. Barium occurs either as an addition to, or else enters the lattice of K-Na-feldspar, plagioclase, muscovite, and biotite, in amounts sufficient for an ore mineralization, and is precipitated as barite, in the hydrothermal phase.

It appears, then, that the minor intrusions, especially their acid phase, are "contaminated" by copper (to a slight extent), lead (more), and barium (much), in the amounts exceeding those for an ore-free mineralization. This suggests a potential ore content in this intrusion. Developed during the hydrothermal stage of the intrusion, was a paragenesis of galena, chalcopyrite, and barite, corresponding to that of lead, copper, and barium, dispersed in the igneous formations.

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The Rostov-on-Don
State University

Received October 13, 1958

POTASSIUM METASOMATISM IN GRANITES OF SOUTHEASTERN TUVA[1]

by

V. P. Yeremeyev

The southeastern part of Tuva, made up chiefly of Proterozoic gneiss and marbles, is cut in a number of places by intrusions of various composition and age, including Precambrian biotite granites, Lower Cambrian ultrabasic rocks and the "Tannuol" Salair complex, and by minor hybrid granite intrusions. In addition, pegmatites are exposed in several places.

Red granites are very distinctive among the assorted "Tannuol" granitoids and are fairly well developed throughout Tuva, especially in its southeast where they occur usually alongside deep-seated faults. We have observed them in the Agar-Dag-Taiga Range area, on the left bank of Tes-Khem River, on top of Ulor Mountain and elsewhere. In central Tuva, they were observed near Balgazik and Vladimirovka villages.

The red granites of different intrusions are very similar, externally; under the microscope, however, they show a difference which has has aroused our curiosity as to their origin.

1. Granites near the Agar-Dag-Taiga Range

Red granites of the Agar-Dag-Taiga Range occur in its southeastern part, in contact with strongly dynamometamorphosed Cambrian schist (Fig. 1). A study of this intrusion has revealed that its coloring is less intense near the peripheral parts (contacts), changing from vivid red to light pink. The petrographic composition of the intrusion, too, is inconsistent: its center consists of medium-grained granite, with fine-grained plagioclase granite about it.

Typical of the central part is specimen No. 263. Macroscopically, this is a medium-grained, fairly fresh rock, sealing-wax red in color, with the crystals about 5 millimeters, on the average. Grayish crystals of quartz with occasional scales of altered biotite are fairly conspicuous on this background. Most conspicuous under the microscope is intense pelitization of the feldspathic fraction, especially affecting the slightly latticed microcline, and to a smaller extent the plagioclase. The latter is represented by orthoclase No. 25, whose edge exhibit a transparent fringe of albite No. 5 ($\gamma' = 1.537 \pm 0.002$; $\alpha = 1.530 \pm 0.002$), while the center is slightly sericitic.

Predominant among the feldspars is a slightly latticed microcline ($\gamma' = 1.529 \pm 0.002$; $\alpha' = 1.523 \pm 0.003$); it is significant that it replaces the plagioclase. As a rule, there is no sharp difference between the crystals of plagioclase and of the slightly latticed microcline. Coarse microcline grains commonly contain relicts of replaced and resorbed plagioclase, and agglomerations of fully chloritized biotite. The interstices between feldspars are occupied by quartz with a typical wavy extinction, and commonly contain resorbed microcline crystals. Locally, quartz penetrates microcline and acquires a granophyric texture. Many magnetite and epidote inclusions are present in chlorite developed on biotite. There are also isolated grains of zircon, titanite, and apatite.

The quantitative mineralogic computation (average for three thin sections) for a specimen is given in Table 1.

The same specimen was chemically analysed (by M.G. Zamurayeva) at the laboratory of the Institute of Ore Deposits, Petrography, Mineralogy, and Geochemistry, Academy of Sciences, U.S.S.R.; the results are given in Table 2, along with analyses of the "Tannuol" gray granite from near Sosnovka village, and of the Elsin River leucocratic granite. It is of interest that the K_2O content, as determined, is smaller in the Agar-Dag-Taiga Range than that anticipated from the microcline content in the rocks (43.3%). This is explained by the lack of uniformity in microcline, with its considerable content of small

[1] O kaliyevom metasomatoze v granitakh yugo-vostochnoy Tuvy.

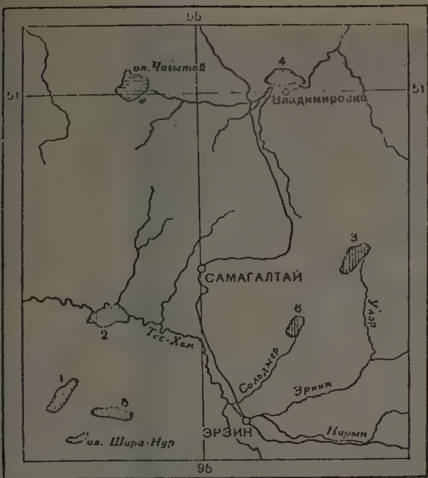

FIGURE 1. Granite outcrops in southeastern Tuva.

1 -- at the Agar-Dag-Taiga Range; 2 -- on Tes-Khem River; 3 -- in the upper course of Ulor River; 4 -- at Vladimirovka village; 5 -- at the Yamalyg Range; 6 -- granitoid gneisses in the upper course of Sol'dzher River.

inclusions of plagioclase and with a slight development of perthite growths. This lack of correspondence was noted and explained earlier, by G.D. Afanas'yev [3].

In the fine-grained peripheral part of the intrusion (specimen 261), unlike its central part, the fresh aspect of the rock is conspicuous, especially in microcline. Here, too, the plagioclase is zoned, being represented by oligoclase No. 26 to 28, in the centers of the crystals, where sericitization and sossuritization are developed. In the peripheral parts, plagioclase is fresh, being albite No. 5 to 7. It is also significant that plagioclase is present here in a considerably larger amount than in the preceding specimen (see Table 1), with its replacement by fresh latticed microcline clearly noticeable. Here, too, the latter is seen to carry relicts of a slightly sericitic and osssuritic plagioclase. The colored and accessory minerals are the same but present in a greater amount.

Rock from the immediate contact with enclosing Cambrian schists (specimen 257) is shattered and represented by plagioclase granite. It is made up almost exclusively of plagioclase, with oligoclase No. 26 to 28 in the center and albite No. 5 in the periphery ($\gamma = 1.533 \pm 0.002$; $\alpha = 1.529 \pm 0.002$). Quartz and microcline occur in very small amounts (see Table 1). The edges of plagioclase crystals are locally replaced by microcline in thin fringes and in larger areas. Present in the interstices are crystals of titanite and apatite and small grains of zircon.

On the whole, the intrusive rock appears to be a plagio-granite, with plagioclase replaced to a great extent by microcline.

2. Granite From the Left Bank of the Tes-Khem

Approximately 30 kilometers northwest of these granites, there are excellent exposures

Table 1

Quantitative Mineralogic Computation for Granites of the Agar-Dag-Taiga Range (in volume %)

Minerals	Central part of intrusion, specimen 263 (average of 3 analyses)	Peripheral part of intrusion, specimen 261, (average of 2 analyses)	Near contact, specimen 257
Plagioclase	11,0	41,0	75,1
Microcline	43,3	23,4	3,8
Quartz	41,9	30,0	16,2
Chlorite and epidote	2,5	4,1	2,7
Titanite	0,2	0,7	1,6
Magnetite	1,1	0,8	0,6
Total	100,0	100,0	100,0

NOTE: Comma represents decimal point.

Table 2
Chemical Analyses of the Tannuol Granites

Components	Red		Gray granite, at Sosnovka village, specimen 58	Microcline, specimen 107, Erzin area
	Agar-Dag-Taiga, specimen 263	Vladimirovka, specimen 324		
SiO_2	74,42	70,01	69,17	75,38
TiO_2	0,16	0,30	0,76	0,19
Al_2O_3	13,55	12,98	13,17	12,83
Fe_2O_3	1,81	4,20	3,09	0,31
FeO	0,25	1,12	2,18	0,95
MnO	0,04	0,06	0,11	0,01
MgO	0,27	0,84	1,62	0,72
CaO	1,23	1,50	3,16	0,59
K_2O	4,42	2,76	2,66	6,41
Na_2O	3,53	4,57	3,10	2,48
H_2O^+	0,32	0,39	0,88	0,54
H_2O^-	0,17	1,47	0,16	0,12
P_2O_5	—	—	—	0,07
S	—	—	—	0,02
Total	100,17	100,20	100,06	100,62
Analyst	M.G. Zamurayeva	O.P. Ostrogorskaya		D.N. Knyazeva

NOTE: Comma represents decimal point.

of red granite on the left bank of Tes-Khem River; they are traceable for about a kilometer along the river, exposed by erosion of the river (Fig. 1). Externally, these granites are similar to the Agar-Dag-Taiga; we believe that both belong to a single eroded and leveled intrusion which forms a flat surface overgrown with grass. In the south (at the Agar-Dag-Taiga Range), granites have been preserved only near the contact with more stable peridotite and serpentine. The enclosing rocks here are Lower Cambrian hybrid hornblende diorite and and gabbro-diorite.

The microscopic study of rocks from this intrusion, too, has shown their similarity, with the central part consisting of granites which change to plagioclase granite toward the periphery. Granite from the central part is strongly altered: plagioclase No. 25 to 27 is sericitized and locally sossuritized. The same picture of microcline replacing plagioclase prevails here, although to a smaller extent than in the Agar-Dag-Taiga Range, with the replacement starting at the edges and proceeding toward the centers of the crystals. The transition boundaries are usually not sharp.

The K-feldspar is also strongly pelitic, with the microcline lattice very poorly expressed. Quartz is present among the feldspars, commonly resorbing the latter and locally forming segments with a granophyric structure.

Colored minerals are fairly abundant, being represented by elongated crystals of hornblende ($c\gamma = 20°$; $2V = 50°$) and by chlorite developed on it. Epidote is much less common. For accessory minerals, there are zircon and apatite. Near contacts, plagioclase granites (specimen 100) are marked by their fresh aspect and by the total lack of K-feldspar. In the rest, they are fairly close to granite.

3. The Ulor Granites

The replacement of plagioclase by microcline is even more obvious in granites exposed along the upper Ulor, in southeastern Tuva (Fig. 2). Here, the granites are not as pelitic as in the central part of the Agar-Dag Range and along Tes-Khem River, although their plagioclase still is strongly sericitic. Relicts of sericitic microcline also persist within large microcline bodies, in addition to fairly coarse scales of secondary muscovite. Other than that, this granite is fairly similar to the Agar-Dag-Taiga varieties.

Also similar to them are granites occurring in central Tuva, at Balgazik and Vladimirovka

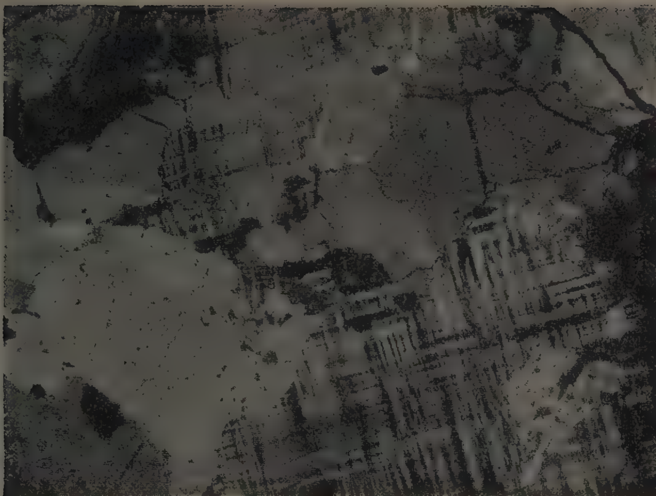

FIGURE 2. Granite from the upper course of Ulor River
Plagioclase relicts in latticed microcline. Specimen 42.

Magnification 30X; Nicols crossed.

villages, and elsewhere. The Vladimirovka granite differs only in its typical granophyric texture and in the less intensive replacement of plagioclase by microcline (Fig. 3).

4. Granites of the Yamalyg Range

Very interesting are granites located 15 kilometers east of the Agar-Dag-Taiga, away from the fault zone. They occur in remnants, on a plain north of Lake Shara-Nur, and form the Yamalyg Range tranding latitudinally for 5 kilometers (Fig. 4). Externally, they are quite different from those described above. They are light-gray gocks with light-pink quartz crystals standing out on the light-colored background of feldspar. Muscovite scales have been observed locally.

A microscopic study (specimen 310) reveals the presence of a large amount of quartz with resorbed crystals of microcline and plagioclase. Small plagioclase crystals, concentrated in some segments of the specimen, alternate with coarser segments of a secondary latticed microcline, to form a porphyritic texture. Locally, plagioclase persists in microcline in small allotriomorphic resorbed relicts. It is very interesting that the microcline crystals, despite their secondary origin, are strongly pelitic. In addition, fairly coarse scales of secondary muscovite are developed on plagioclase.

It appears that this microclinization of the Tuva granites is related to the latest post-igneous alkaline solutions which, in their reaction with plagioclase granites, were the cause of a metasomatic replacement of plagioclase by microcline. The source of these solutions was the same magmatic hearth from which the granites themselves originated. The zonal effect of these solutions on the intrusive body of the Agar-Dag-Taiga Range is very typical, and is confirmed by the less intensive microclinization in its peripheral parts.

The occurrence of red granite in the vicinity of deep faults might suggest the deep faults were vents for the alkaline solutions. However, the presence of K-metasomatism in the Yamalyg Range granites, away from the faults, suggests rather that these solutions came from the depths of intrusive hearths, after the intrusions had been consolidated, and altered them autometasomatically. Judging from the fact that not only plagioclase but the secondary latticed microcline, too, is usually resorbed in the quartz of plagioclase granite, it may be assumed that microclinization, or the K-metasomatism, preceded an acid stage of the post-igneous process. These data are in full accord with the D.S. Korzhinsky [6] theory of metasomatic processes.

Metasomatism of the Tuva plagioclase granites shows considerable similarity with processes described earlier by G.D. Afanas'yev

FIGURE 3. Granite at Vladimirovka. Granophyric growth of quartz in orthoclase, about plagioclase.

Specimen 324. Magnification 45X; Nicols crossed.

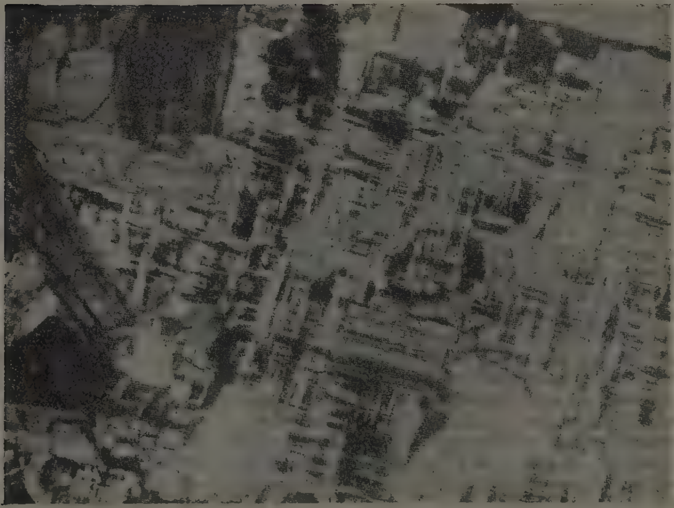

FIGURE 4. Granite from the Yamalyg Range. Small plagioclase crystals are replaced by coarse porphyritic "incrustations" of microcline.

Specimen 310; magnification 30X; Nicols crossed.

[1, 2] in connection with Paleozoic granodiorite of the Main Caucasian Range. He also notes that the fringe diorite in the Malolabinsk massif changes gradually to granite of the central part. He points out the two different formational stages for massifs. Stage one, the intrusion of a granodiorite magma and the formation of rocks ranging from diorite to granodiorite; stage two, an alternation of the already consolidated intrusives to microcline granite of the central parts and to granodiorite of the periphery, with the microclinization gradually decreasing in intensity in that direction.

This phenomenon is fully corroborated by D.S. Korzhinsky's views [6] on post-igneous metasomatism wherein a partial post-igneous replacement of acid plagioclase by either orthoclase or microcline takes place in granite and syenite.

The K-metasomatism processes are fairly common in granites of central and southeastern Tuva. They have been observed also in granitoid gneisses on the left bank of Balgashta River (tributary of Sol'dzher River) which we believe to have been formed as a result of the intrusion of a granite magma into gneiss, with the gneissoid texture preserved in their apical parts.

Macroscopically, there is a semblance of relicts of coarse yellowish plagioclase crystals standing out against a gray micaceous groundmass. Under the microscope, even this proves to be a metasomatic replacement of plagioclase — by fresh latticed microcline, and by muscovite either in fine laminae or coarse plates. Quartz is concentrated in segments of coarse grains with resorbed plagioclase grains. There is a considerable amount of fairly coarse crystals of apatite, and less commonly of zircon, sphene, and magnetite, all as a result of assimilation of the enclosing rocks.

CONCLUSION

A petrographic study of the southeastern Tuva granites shows an intrusion and consolidation of post-Cambrian, pre-Devonian plagioclase granites. Subsequently, they were acted upon by alkaline solutions which we, like the earlier students (G.D. Afanas'yev, D.S. Korzhinskiy, and others) associate with with the intrusions themselves. As a result, the intrusions were autometasomatically altered, which has led to the replacement of plagioclase by latticed microcline. The intrusion as a whole is marked by a definite zonation, because the replacement process was most complete in its central part, and less so toward the periphery.

This replacement phenomenon, described from Tuva, corroborates the earlier assumption by G.D. Afanas'yev [1] that the microclinization process is widely developed not only in the caucasian but also in other granodiorite intrusive massifs.

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Institute of Geology of Ore Deposits,
Petrography, Mineralogy, and Geochemistry,
Academy of Sciences of the U.S.S.R.
Moscow

Received February 24, 1959

MICA FLUORITE TOPAZ MINERALIZATION SUPERIMPOSED ON SULFIDE MINERALIZATION IN THE SHCHERBAKOV ORE FIELD[1]

by

B. A. Radkevich and I. M. Tomson

This paper describes a peculiar mica-fluorite mineral association superimposed on sulfide ores. The appearance of this later and high-temperature association with topaz, muscovite, and cassiterite suggests a rising temperature of solutions in the process of mineral formation, and is related to a rejuvenation of the deep magmatic source.

* * * * *

The Shcherbakov ore field, represented by tin-polymetal cassiterite-sulfide veins, has been found to have been affected by a high-temperature greisen-like mica-fluorite-topaz mineralization, with cassiterite, which is younger than the veins and which is superimposed on the polymetal mineralization. The mica-fluorite-topaz veins are best developed in the Levoberezhnyy deposit where they have recently been uncovered by underground mining works. It is significant that the mica-fluorite-topaz association is missing in almost all other cassiterite-sulfide deposits in the Maritime Province related to the Shcherbakov ore field. The only exception is the Stalin deposit where small accumulations of a light-colored mica have been observed in association with an earlier quartz.

The Shcherbakov ore center in the South Maritime Province is associated with a brachianticlinal uplift in Upper Cretaceous extrusives overlying upper Paleozoic (?), Triassic (?), and Lower Cretaceous sedimentary rocks.

Ore veins of this deposit cut chiefly sedimentary rocks — sandstone and shale, mostly of Paleozoic age. They form a series of vein zones associated with parallel faults trending northwest and outline a major regional zone of weakness which trends northwest along the Pkhusun Valley, from the Sea of Japan to the Sikhote-Alin crest. The ore center is confined to where this tectonic zone cuts sedimentary rocks exposed from under the Ol'ginsk volcanics.

Granitoid massifs are not known from the Shcherbakov area. Only dikes of porphyrites and quartz porphyries are present here, locally accompanied by disseminated mineralization.

The Shcherbakov vein series is associated with two principal parallel fault zones trending northwest. Associated with the southwestern zone are the Shcherbakov (Shakhtnyy district) and Sobach'ya Pad'; the Silinsk and Levoberezhnyy deposits are associated with the northern zone.

Although individual districts of the Shcherbakov ore field are located along the same tectonic zones, they differ in the nature of their mineralization. Most typical of this area are cassiterite-sulfide veins accompanied by ferro-manganese carbonates and locally by quartz. These are multistage veins, a product of several consecutive paragenetic mineral associations separated by shattering. The contact alterations which accompany these veins are inconspicuous, being expressed by a slight chloritization.

Present in the ore field are the following paragenetic associations corresponding to different stages of mineralization:

1. Quartz-arsenopyrite with cassiterite (Shcherbakov deposit);

2. Sulfide, pyrrhotite-galena-sphalerite association with cassiterite. This is the main stage, common to all deposits of the Shcherbakov center;

3. Carbonate-sulfide association with gray copper ores, boulangerite, stannite, pyrrhotite, sphalerite, and galena. Carbonates are represented by ankerite, manganese-ankerite,

[1] Nalozheniye slyudisto-flyuoritovo-topazovoy mineralizatsii na sul'fidnuyu mineralizatsiyu v Shcherbakovskom rudnom pole.

and ferridolomite.

4. Chalcedony-calcite-antimony association with pyrrhotite, boulangerite, jamesonite, antimonite, and other antimony minerals. This terminal stage is expressed in the Silinsk deposit in independent veins, also in isolated younger vein bands localized near the walls of sulfide-carbonate veins.

All these mineral associations are, on the whole, intermediate- to low-temperature.

Typical of the ores are relicts of collomorphic textures and structures which suggest the important part played by colloids in the ore-forming process; also the diversified banded and "stream-lined" textures, a result of the ore metamorphism related to recurring small shifts along the vein cracks.

These mineral associations were formed as consecutive stages of a discrete process; this is substantiated by their spatial coincidence and by the recurrent appearance of the same minerals.

The appearance of different stages in mineral formation is determined by recurrent reopening of vein fractures, in the mineralization process from solutions of a variable composition, and with an ever-falling temperature. The terminal chalcedony-antimonite stage took place under low-temperature conditions. It is possible, however, that the preceding carbonate-sulfide ores, too, were formed at fairly low temperatures.

The discovery of mica-fluorite-topaz mineral segregations, with cassiterite, cutting the sulfide veins in the Shcherbakov ores, was a surprise. Such mica-fluorite formations were previously known from the shaft district of the Shcherbakov deposits; they have recently been discovered also in the Levoberezhnyy deposit. Surprisingly enough, these mica-fluorite-topaz aggregates definitely cut the earlier sphalerite-pyrrhotite veins. Clearly exposed in a shaft are vein bands of a light-colored micaceous rock, up to 20 centimeters thick, running along the wall of a dark sphalerite or massive pyrrhotite vein, locally crossing the latter (Fig. 1). These broad micaceous bands send off thin veinlets into the sulfide body, where they wedge out and disappear. The radial and coarse-grained aggregates of silvery mica carry distinct xenoliths of massive sulfide ores, as peculiar breccia-like inclusions (Fig. 2).

Principal minerals in the mica-fluorite-topaz aggregate, in this younger vein groundmass cutting the sulfide ores, are mica and fluorite; topaz is not as well developed and considerably altered by secondary products. Also present is quartz, very unevenly

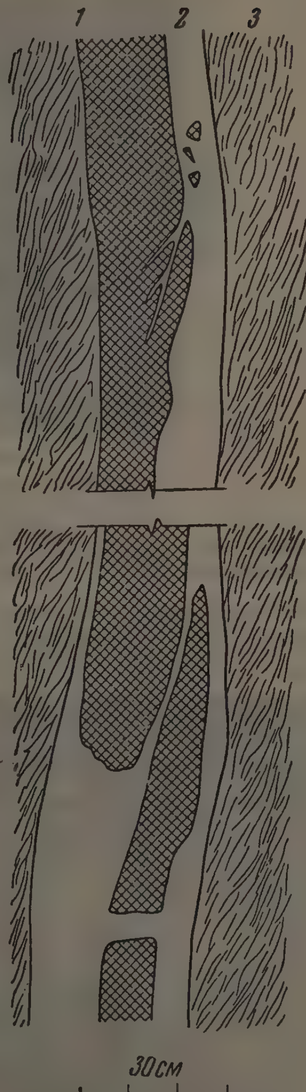

FIGURE 1. A mica-fluorite-topaz vein crossing the near-wall zones of a sulfide pyrrhotite-sphalerite vein.

1 -- mica-fluorite-topaz vein; 2 -- sulfide pyrrhotite-sphalerite vein; 3 -- enclosing schist (a sketch of the mine drift roof).

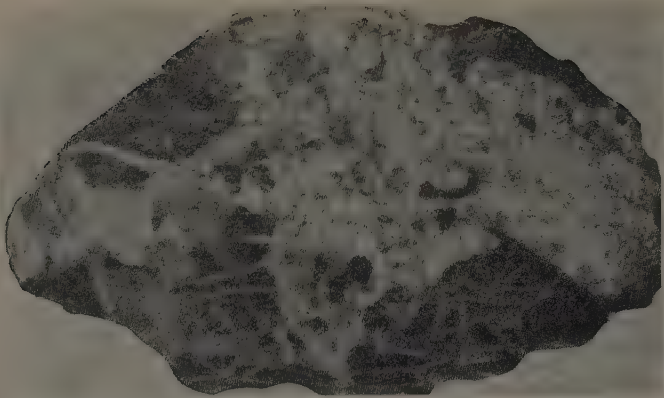

FIGURE 2. A mica-fluorite association (light-colored) cutting and cementing earlier sphalerite and pyrrhotite (gray and dark).

Polished section. Natural size.

distributed, and a small amount of cassiterite, arsenopyrite, etc.

The textural pattern of these ores is fairly diversified, with distinct segments of coarse-scaled mica aggregates in rosettes, radial clusters, and matted aggregates, also of a dense, fine-grained mass representing a pseudomorph on some tabular and columnar minerals (Fig. 3) where mica is in a close association with fluorite. These mica-fluorite aggregates commonly carry pseudocollomorphic, parallel, festooned to wavy bands and rosettes. The fine-grained mica aggregates are made up of reniform bodies. Topaz occurs in isolated bands in the mica-fluorite association, also forming finely-crystalline acicular radial structures.

The general textural pattern of these formations is determined by a wide development of radial banded, festooned, and in places dense finely crystalline aggregates. These common textural features seem to suggest a rapid crystallization at a rapidly falling temperature.

A microscopic study of the mica-fluorite-topaz aggregates has revealed a great variety of microtextures within them. Locally, mica is developed within distinct crystallographic outlines of an undeterminable original mineral, where it is disposed in typical parallel bands probably corresponding to the cleavage planes

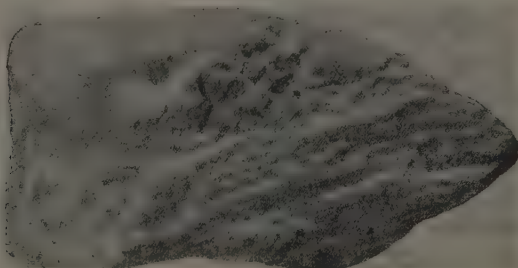

FIGURE 3. Pseudomorph of fine-grained mica aggregate on a fully replaced columnar mineral.

Specimen. Natural size.

FIGURE 4. Crystal of an undetermined mineral, replaced by mica (in the lower part of picture), with edges showing a chain of cassiterite grains (dark). A zone of radial topaz in the upper part of the picture. Thin section No. 18.

Magnification 27X; Nicols crossed.

of that mineral. Individual micas, in their turn, grow on these former cleavage planes, in transversely-oriented rosettes. It is quite possible that we deal here with fully replaced crystals of feldspar or topaz (Fig. 4). The faces of replaced crystals show chains of cassiterite grains and fringes of fine-scaled mica with rosettes of finely-radial topaz forming on it fan-shaped aggregates of fine needles. These topaz rosettes exhibit a wavy extinction. Fluorite participates in some pseudomorphs along with mica; a system of oblique fluorite bands emphasizes the direction of cleavage in the original mineral (Fig. 5).

Reactive relations between minerals of the mica-fluorite-topaz association are very common; in addition to the examples mentioned above, they are expressed in the replacement of topaz by mica and in the penetration of mica along the cleavage planes of ancestral fluorite crystals. Also present are well developed depositional structures. For instance, spherulitic aggregates and rosettes of coarse-scaled mica with a well-defined concentrically-zoned structure (Fig. 6) were formed evidently in already open hollows, as a result of their rapid crystallization out of solutions.

Associated with mica is cassiterite which also occurs in rosettes and radial segregates of acicular crystals (Fig. 7). Elsewhere,

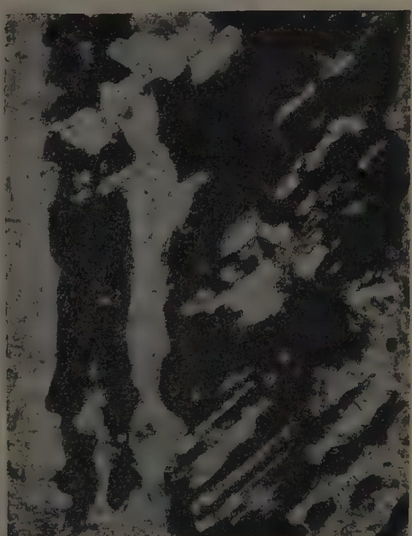

FIGURE 5. Fluorite (black) and mica, forming a pseudomorph on an undetermined mineral. Thin slide No. 33.

Magnification 24X; Nicols crossed.

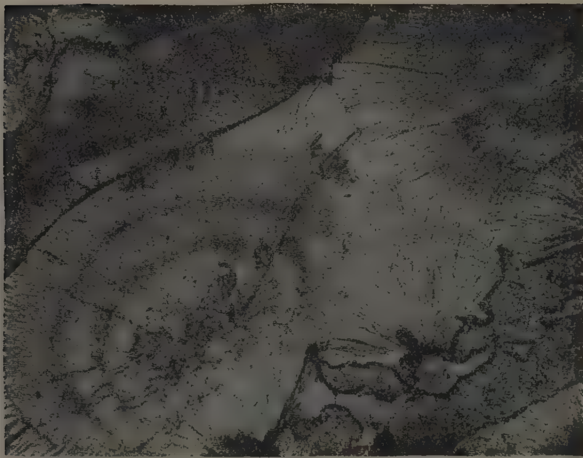

FIGURE 6. Concentrically zoned spheres of mica.
Thin section No. 33.

Magnification 27X; Nicols crossed.

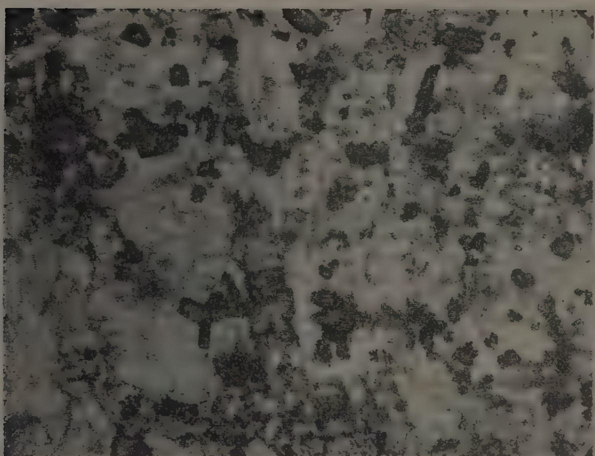

FIGURE 7. Cassiterite crystals in a mica aggregate.
Thin Section No. 29.

Magnification 63X; Nicols crossed.

cassiterite forms finest granules and growths in a mica-fluorite aggregate — in twins and triplets dispersed in the micaceous groundmass. The Levoberezhnyy cassiterite deposit is related on the whole to this mica-fluorite-topaz vein body.

Given below is a brief description of some features of principal minerals in the mica-fluorite-topaz association. The light-colored mica of this association has been studied microscopically and by X-rays; chemically, it has been identified as muscovite with a

Table 1

Chemical Analysis of Mica[a]

Oxides	Weight %	Weight %, without CaF_2	Molecular number
Si_2	47,45	48,57	809
TiO_2	traces		
Al_2O_3	33,44	34,23	336
Fe_2O_3	not det.		
FeO	1,74	1,78	025
MnO	0,09	0,09	001
MgO	0,64	0,66	016
CaO	1,76	—	
Na_2O	0,22	0,22	003
K_2O	9,50	9,72	103
H_2O^-	0,12		
H_2O^+	4,62	4,73	261
F	1,15		
Total	100,73		
$O=F_2$	0,48		
	100,25		

[a]The analysis of this mica was computed from the following formula: $K_{0.8}$ (Fe, Mg)$_{0.1}$ $Al_{1.9}(Al_{0.8}$ $Si_{3.2}O_{10})$ $(OH)_2 0.05H_2O$. Thus, in its composition, this mica corresponds to alumophengite, close to common muscovite.

The analysis was performed in the central chemical laboratory of the Institute of Geology of Ore Deposits, Petrography, Mineralogy, and Geochemistry, Academy of Sciences, U.S.S.R. Ye.I. Lomeyko, Analyst.

NOTE: Comma represents decimal point.

higher-than-average alumina and iron content (see Table 1).

Quartz is present in various amounts, commonly with crystalline faces showing a multitude of gaseous-liquid inclusions and small specks of mica.

Fluorite was deposited during the entire period of formation of this association. It forms early crystalline inclusions of a distinct form; along with mica, it participates in the building of comparatively late fine-grained aggregates.

Topaz forms radial clusters of acicular crystals. Its refractive indices, measured in an immersion fluid, were as follows: $\gamma = 1.1616$, $\alpha = 1.605$.

Cassiterite occurs in acicular and prismatic segregations, in places in complex growths. It occurs occasionally in collomorphic aggregates.

As already noted, the mica-fluorite association was formed after an intensive shattering of the pyrrhotite-galena-sphalerite association; thin veinlets of this younger mineral formation occasionally penetrate the microfaults which cut and displace the twins of coarsely crystalline sphalerite. It appears, however, that some sulfides continued to be deposited along with the micaceous aggregate; specifically, chains of coarsely crystalline galena aggregates have been observed at times within the wavy bands of a mica-fluorite vein body. Locally, arsenopyrite incrustations are associated with the sulfide aggregates. Finally, in a number of places, the micaceous bands are cut by veinlets of sulfides, sphalerite, chalcopyrite, and fine-grained galena; these sulfide veinlets, going along their trend, change to veinlets of purple fluorite (Fig. 8). Consequently, the formation of mica-fluorite-vein aggregates was followed by a new sulfide stage. Thus the mica-fluorite-topaz mineralization found itself in a "fork" of two subsequent sulfide stages.

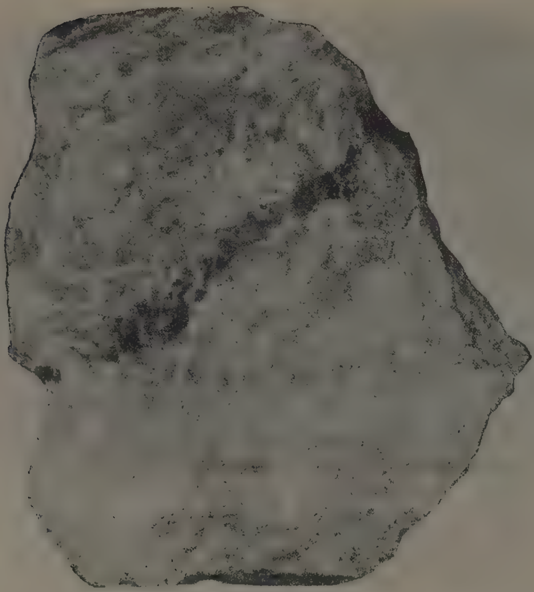

FIGURE 8. A sulfide veinlet cutting the mica-fluorite association.

One-half natural size.

Of particular interest is the fact that the paragenetic association of mica-fluorite-topaz veins is similar to that of high-temperature greisens. However, the formation conditions of the Shcherbakovka micaceous veins are quite different from those of greisens. Being formed between the two stages of sulfide mineralization, the veins occur in schist, away from granitoid intrusives and even outside the hornfels halo. Present near the deposit is only a small neck of quartz porphyry which appears to be accompanied by a dispersion sulfide mineralization.

The wide development of radial structures, and the concentric and festoon-like distribution of mineral bands point to a rapid crystallization of minerals in this association.

The above-mentioned features of the mica-fluorite-topaz association, and the extensive development of reactions between its minerals, are determined by its formation in a radically changing physical and chemical environment. The appearance of a mica-fluorite-topaz mineraliation probably was accompanied by a higher temperature, compared with the preceding sulfide stage.

In the type of mineralization, this association differs substantially from cassiterite sulfide ores common in the eastern Maritime Province. The association of these ores with similar mica-fluorite-topaz formations with cassiterite, more typical of the cassiterite-quartz formation, shows that a sharp boundary cannot always be drawn between the cassiterite-sulfide and cassiterite-quartz deposits, and that a superposition of one type over the other is possible. In our area, these micaceous formations came after, probably a result of the rejuvenation brought about by an additional injection of magma, after the formation of the main body of comparatively low-temperature cassiterite-sulfide ore veins. It is not impossible that the presence of such diversified but contemporaneous cassiterite-sulfide and mica-fluorite-topaz mineralizations is related to sources lying at different depths.

Institute of Geology of Ore Deposits,
Petrography, Mineralogy, and Geochemistry,
Academy of Sciences, U.S.S.R.
Moscow

Received March 16, 1958

BRIEF COMMUNICATIONS

NEYVITE — A NEW VEIN ROCK[1]

by

N.D. Sobolev

Ultrabasic rocks of the Urals belong, according to A.N. Zavaritskiy [3] and G.L. Padalka [6], to a single extrusive complex, rich in various ultrabasic to acid vein rocks. According to P.M. Tatarinov [7], the Uralian ultrabasics form three belts: the western, central, and eastern. In the central belt, the largest massif trending meridionally (from Nev'yansk to Verkhotur'ye), is made up of peridotite and serpentine; it was studied by the author in the Neyva-Tagil watershed. This massif is cut by numerous veins designated as microdiorite and diorite-porphyrite, by V.V. Arshinov, B.Ya. Merenkov [1], and A.Ye. Malakhov [5]. Students of the western and eastern zones believe these veins to be co-magmatic with the ultrabasic intrusion and to belong to a vein series from dunite to albitite, plagioaplite, and aplite.

Very similar to these veins are those in the west belt, described by N.K. Vysotskiy [2] as albite amphibolite and hornblende aplite. That author notes that the albite amphibolite is usually schistose but carries massive veins clearly cutting the basic rocks and suggesting their igneous origin. With regard to microciorite, V.V. Arshinov and B.Ya. Merenkov note that "the amount of hornblende increases to such an extent, and the rectilinear orientation of its crystals is so evident, that the rock can be easily taken for amphibolite." ([1], p. 20.) The diagram in Figure 1 describes these rocks as rich in alkaline aluminosilicates with a definite predominance of sodium, i.e., of albite. In this type of rock, the factor c does not reflect the relative amount of anorthite but rather of hornblende with up to 12% of calcium oxide and up to 20% of aluminum oxide — which is responsible for the value of factor c.

It appears from the results of 18 chemical analyses of amphibolite, microdiorite, and albite of the Urals (Fig. 1), designated by the authors as albite-hornblende rocks, that the nearer a point is to axis S (left part of diagram), the greater is the albite in the composition of the rock (the nearest point is 3-27, albite described by A.N. Zavaritskiy, [3]); the farther away from the axis, the greater the importance of hornblende (extreme points V-508 and V-509 are albite amphibolite described by N.K. Vysotskiy). According to A.N. Zavaritskiy, such albite amphibolites and standard amphibolites should be designated as hornblendite ([4], p. 210). Most of the intermediate points between them and albitite are described by him as albite-hornblende rocks, under the name microdiorites.

Transitions between microciorite and hornblendite, between microdiorite and albitite, and betweel albitite and aplite (hornblende aplite of N.K. Vysotskiy) have been noted by many students of ultrabasic massifs of the Urals, including this author.

Some investigators believe that microdiorite is of an epi-magmatic origin, with hornblende formed in place of pyroxene, and albite in place of basic or intermediate plagioclase; some others believe that they are vein offshoots from diorite massifs and that they carry plagioclase No. 33 to 41 (A.Ye. Malakhov);[2] finally, there are some who regard microdiorite as lamprophyre in which either the basic or the intermediate plagioclase has been albitized. As we see it, all those stu dents did not have at their disposal material necessary for a solution of the problem of the origin of these veins. Our own data, obtained from the study of quarry specimens and of cuttings and cores from boreholes, shed more light on the subject. These vein rocks are greenish-gray, massive, consisting of approximately the same amount of hornblende and

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[1] Neyvit—novaya gornaya poroda iz gruppy zhil'nykh.

[2] The plagioclase in veins is Albite No. 5. Obviously, A.Ye. Malakhov did not check his identification of plagioclase by the Fedorov method from refractive index determination; hence this error.

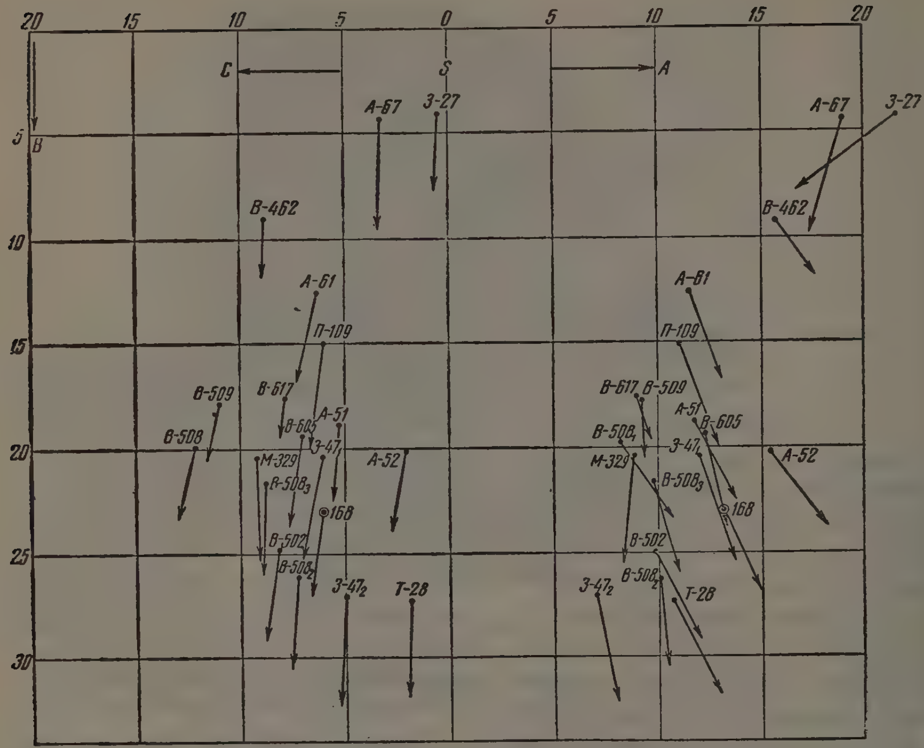

FIGURE 1. Diagram after A.N. Zavaritskiy

Each vector is marked by the initial letter of the author's name and the page number of the work cited.

albite, with a grain size of 0.5 to 1 millimeter. The hornblende most commonly has idiomorphic outlines, with regard to albite, while the rock is hypidiomorphic, less commonly panidiomorphic.

Occasionally, albite has an incipient zoned structure, with the central part replaced to a small extent by fine-grained zoisite; the latter is occasionally developed in non-zoned grains, as well. This plagioclase has been determined by the Fedorov and the immersion methods as No. 1 to 6, with Carlsbad, less commonly albite and Baveno twinning. According to the chemical analysis, corroborated by the Fedorov and the immersion methods, the central part of the albite grains consists of plagioclase No. 6; the peripheral part, of No. 1. The refractive indices, determined by the parallel light beam method, are as follows: $\gamma = 1.5395$; $\beta = 1.5285$; $\gamma - \beta = 0.011$.

The color of hornblende is somewhat anomalous, beige to greenish; the absorption, $\gamma > \beta > \alpha$; optical properties: $\gamma - \alpha = 0.021$; $\gamma - \beta = 0.011$; $\beta - \alpha = 0.010$; $2V = -82°$; $c\gamma = 23°$. As a rule, hornblende is replaced by a fine-grained aggregate of zoisite, in places with an addition of epidote and chlorite. No evidence of uralitization of hornblende has been observed. Present among accessory minerals are occasional fine grains of magnetite and very rarely of sphene and apatite.

Cores from boreholes contain specimens without secondary minerals. One of them, No. 168 from 50 meters deep, was differentiated into monomineral fractions whose chemical analyses, along with the overall analysis, are given in Table 1. For control, there was the hornblende fraction from specimen 483 (from a depth of 32 meters, in another borehole). The results of the chemical analysis are given in the same table.

According to a chemical analysis of the

Table 1

Oxides	Specimen 168, overall	Specimen 168, albite	Specimen 168, hornblende	Specimen 483, hornblende
SiO_2	47,51	62,54	41,88	41,65
TiO_2	0,72	0,09	1,05	0,96
Al_2O_3	17,76	18,99	15,82	15,83
Fe_2O_3	2,79	0,79	3,13	4,25
FeO	7,62	1,08	10,32	8,90
MgO	4,80	0,60	12,83	13,75
MnO	0,16	—	—	—
NiO	0,05	—	—	—
CaO	7,96	2,62	11,04	11,26
Na_2O	4,91	10,02	2,44	2,34
K_2O *	1,12	0,38	0,09	0,10
P_2O_5	0,20	0,09	—	—
H_2O^+	2,59	—	1,40	1,07
H_2O^-	0,27	0,16	0,00	0,00
CO_2	1,64	—	—	—
S	0,03	—	—	—
Loss on ignition	—	2,52	—	—
Cr_2O_3	0,001	—	—	—
ZrO_2	0,03	—	—	—
V_2O_5	0,03	—	—	—
CoO	0,00	—	—	—
CuO	0,00	—	—	—
F	—	—	0,22	0,22
Total	100,20	99,88	100,22	100,33

*The amount of potassium is 0.6% more in the overall chemical analysis than in albite and hornblende. It appears that the rock contains undeterminable orthoclase, up to 3%.

NOTE: Comma represents decimal point.

monomineral fraction, the albite is 93% albite No. 7 and 7% an acicular addition of hornblende intergrown with albite.

A computation of both analyses of the monomineral fractions of hornblende shows that they correspond to the following composition:

Specimen 168, $Na_{0.6}Ca_{1.8}Mg_{2.8}\overset{++}{Fe}_{1.3}(Al_{1.0}\overset{+++}{Fe}_{0.3})_{1.3}$
$[(Si_{6.1}Al_{1.8}Ti_{0.1})_8O_{22}][(OH)_{1.4}F_{0.1}O_{0.5}]_2$,

Specimen 483, $Na_{0.7}Ca_{1.7}Mg_{3.0}\overset{++}{Fe}_{1.1}(Al_{1.0}Fe_{0.3})_{1.3}$
$[(Si_{6.1}Al_{1.8}Ti_{0.1})_8O_{22}][(OH)_{1.1}F_{0.1}O_{0.8}]_2$

These hornblendes belong to a somewhat basaltic common variety, which is reflected in their coloring.

A conversion of the overall chemical analysis of specimen 168 to the modal mineral composition, along with a quantitative mineral computation in thin sections, is given in Table 2.

The minerals were isolated in the order indicated in the weight percentage column; residual oxides for hornblende correspond to the following formula for it:

$$Na_{0.6}Ca_{1.5}Mg_{1.9}\overset{++}{Fe}_{0.6}(Al_{0.5}\overset{+++}{Fe}_{0.4})_{0.9}$$
$$[(Si_{5.4}Al_{2.5}Ti_{0.1})_8O_{22}][OH]_2.$$

Many tens of such veins were studied at different depths (down to 100 meters below the surface); the results show that most of them, in the optic properties of their rock-forming minerals and in their quantitative ratios and structure, are close to specimen 168.

In addition to this type of veins, their porphyritic varieties occur in the massif: hornblende predominates over albite in the inclusions, with the grain size up to 3 millimeters, and with 3 to 5 grains per centimeter square (in thin section). In addition, there are non-porphyritic varieties with

Table 2

Minerals	% weight	% volume	without calcite	As computed in thin sections
Calcite*	3,6	3,9	—	—
Sphene	0,6	0,5	0,5	accessory
Apatite	0,3	0,3	0,3	
Pyrite	0,1	0,6	0,6	
Magnetite	1,0			
Albite No. 6	40,0	44,6	46,9	47
Hornblende	54,6	49,2	51,7	52

*In microveins, up to 0.1 millimeter thick.

NOTE: Comma represents decimal point.

dominant hornblende or albite, changing to hornblendite and albitite.

A detailed study of the thickest veins has revealed two phases of intrusion. The second phase is represented by thin veins (up to 10 centimeters) which cut veins of the first phase along their parallelepipedal cleavage planes. In one instance, the second phase rock was a porphyritic variety; in another, a non-porphyritic, with dominant albite. In both, the second phase rocks carried distinct xenoliths (up to 5 centimeters across) of the first phase rocks. At contacts with the second phase veins, hornblende in the first phase rocks has been replaced, slightly to considerably, by zoisite; such hornblende, too, is regarded as xenolithic (Fig. 2).

These facts preclude a post-magmatic origin of the veins, first because of the presence of unaltered rocks free of secondary minerals; second, because of the definite, albeit uncommon, zonation of albite; and third, because of the total lack of relict minerals and structures.

An epi-magmatic origin of these veins would imply a reworking not only of the mineral composition of the rock but of its structure as well. No such thing has been observed; this is especially graphically illustrated by the rhombic sections of hornblende (in inclusions and in groundmass), which are not characteristic of pyroxene.

The decisive argument for an igneous origin of these rocks is the presence of the second phase, in veins cutting those of the first intrusive stage, and the presence of numerous xenoliths of the first phase in the second (Fig. 2). Both in contacts and in the xenoliths, rocks of the first phase have been altered, with their hornblende replaced, occasionally more than half, by zoisite. Should there have been a subsequent albitization, uralitization, or any other "greenstone alteration," the second and the first intrusive phases would have been indistinguishable; as it is, they differ sharply, both in outcrops and in thin section.

The regularity in the formation of such veins, their massive aspect and a definitely magmatic origin, make it expedient to designate these rocks as a new group named neyvite (after the Neyva River). The need for such a designation is determined by the importance of neyvite in the formation of minerals genetically related to it. Their continued designation as albite amphibolite, microdiorite, lamprophyre, etc., tends to complicate and conceal the true problems of the geology of intrusions and of the origin of minerals.

Moreover, the origin of neyvite is often interpreted incorrectly as a result of "greenstone regeneration" of vein rocks. In order fully to clarify this subject, we shall turn to the following two statements of A.N. Zavaritskiy:

1. "Gabbroic pegmatite and plagioaplite are vein rocks similar to those associated with gabbro and representing a residue after the differentiation and crystallization of a gabbro magma. Albitite may be regarded as a result of desilication of plagioclase in a reaction with dunite."

2. "A group different from all of the above-named vein rocks, related to dunite, is represented by greenstone gabbroid rocks

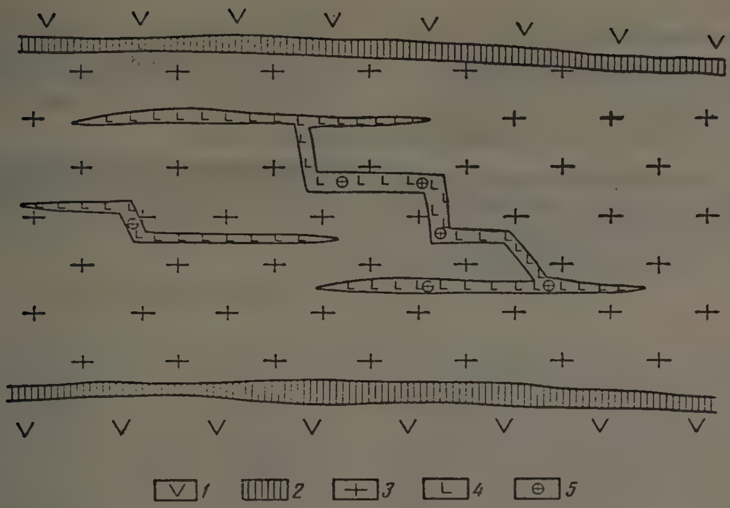

FIGURE 2. Neyvite vein with two stages of magma intrusion. Sketch of a quarry exposure.

1 -- serpentine; 2 -- wall zone chlorite; 3 -- neyvite of the first intrusion phase; 4 -- neyvite of the second intrusion phase; 5 -- xenoliths of the second intrusion phase.

of various grain size (metadiorite, metagabbro, etc.), very common in serpentine massifs. Primary minerals in these rocks were altered, as a rule, in a 'greenstone regeneration,' with plagioclase usually albitized and the colored minerals represented by uralite hornblende" ([4], p. 250).

We reiterate that no evidence of uralitization has been found in neyvite.

In the light of these statements of A.N. Zavaritskiy, albitite should be regarded as an ultimate residue of differentiation and crystallization of a gabbro magma, with its desilication in a reaction with the enclosing rocks. Considering that albitite is related to hornblendite by way of neyvite, this statement of A.N. Zavaritskiy is brought to its logical conclusion by being expanded to embrace the entire range of interactions of the residual melt with the enclosing rocks. The extreme expressions of such a reaction are hornblendite and albitite.

The feasibility of differentiation of a new rock, neyvite, is justified petrographically in nepheline-pyroxene rocks of the series pyroxenite-jakupirangite-melteigite-iolite-urtite, differing in their quantitative ratios of pyroxene and nepheline.

In our example, the differentiation is as follows: hornblendite, with up to 20% albite; neyvite, with 20 to 80%; and albitite, with 80 to 100% albite.

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All-Union Institute of Mineral Resources,
Ministry of Geology and Conservation
of Mineral Resources of the U.S.S.R.
Moscow

Received January 30, 1958

AT THE FOUNTAINHEAD OF CHINESE-SOVIET SCIENTIFIC COLLABORATION[1]

by

L. P. Kletskiy

In the first half of 1958, the Academy of Sciences, U.S.S.R., elected the Chinese geologist Li Sy-Huan as one of its foreign members. One year later, in May, 1959, the Presidium of the Academy awarded Li Sy-Huan the A.P. Karpinskiy gold medal. At about the same time, news came from the Chinese People's Republic that Li Sy-Huan had been appointed acting director of the Central Office of the Society for Chinese-Soviet Friendship.

Each of these two acts is a well-deserved recognition of the great scientific and social activities of one of the leading scientists of a brotherly republic. Professor Li Sy-Huan is not only an outstanding Chinese geologist but a well-known statesman and a champion of long standing of rapprochement with the Soviet Union. He is one of those who has stood at the fountainhead of scientific cooperation between the two countries.

* * *

A new page was turned in the intellectual life of China, after the famous "May the Fourth" (1919) anti-imperialist coup which, in the words of Mao Tse-Tung, has led to the birth of a new culture in China. Being an echo of the Great October Socialist Revolution in Russia, the "Fourth of May" movement has promoted an interest in the Soviet State, its political constitution, economic policy, its social and cultural life, and its science.

Great interest in Soviet Science was evinced also by Chinese geologists. By the fall of 1925, the Geologic Survey of China established relations with ten Soviet scientific organizations, including the Academy of Sciences, Geological Committee of the RSFSR and its branches in Moscow and Tomsk, and with the Mineralogical, Paleontological, Geographic, Natural History, and other societies.

[1] U istokov Kitaysko-Sovetskogo nauchnogo sotrudnichestva.

This intensified interest of Chinese geologists in Soviet science was quite natural. Our country was completing the restoration of its economy and was entering a period of building it up. The geologists were confronted with a responsible task; they have proved themselves equal to it. Applied geology, based on theoretical achievements, has played a great part in the discovery and development of productive potential of this country.

An important event in Chinese-Soviet scientific relations was the participation of a noted Chinese geologist, veteran of the revolutionary movement, Professor Li Sy-Huan, in the celebration of the Bicentenary of the Academy of Sciences, U.S.S.R.

Li Sy-Huan is known as one of the founders of Chinese geologic science. A number of his works were known to Soviet scientists. The Academician A.Ye. Fersman, in working out his idea of geophysical mapping, used, along with other material, the results of a new method of geologic study of intrusions[2] proposed by Li Sy-Huan.

A.Ye. Fersman had learned about this discovery by the Chinese scientist from the Bulletin of the Geological Society of China.[3]

Looking back, Li Sy-Huan has this to say about his association with Soviet geologists: "In the Soviet Union, I had the opportunity to meet many geologists of the older generation, such as the President of the Academy then, A.P. Karpinskiy, and Professor of the Moscow University, A.P. Pavlov. To me, they personified the Soviet scientists ardently concerned in the development of China. I met more often with A.A. Borisyakom, A.Ye. Fersman, and others.

"I discussed the subject of joint work of Soviet and Chinese scientists and methods for

[2] A.Ye. Fersman, Izbrannyye Trudy, t. 3, Izd. Akademiya Nauk SSSR, 1955, p. 633.

[3] 1924, III, 2, pp. 109-115.

such study with some of the Soviet Geologists of about my own age. Their sincere desire for cooperation with Chinese scientists has also made a deep impression on me."[4]

In a conversation with A.N. Krishtofovich, a well-known Soviet paleontologist, Li Sy-Huan extended an invitation to Soviet geologists, to study the Cenozoic phytomorphs of China. The Chinese scientist believed that such a task could be fulfilled only by the cooperation of geologists of the two countries. He discussed with M.A. Bolkhovitinova the differentiation of the Carboniferous in China and the U.S.S.R. Li Sy-Huan had a high opinion of M.A. Bolkhovitina's work in the Carboniferous stratigraphy and was pleased with her readiness to participate in the organization of Sino-Soviet geologic study. Li Sy-Huan came to an understanding with A.S. Borisyak, a Corresponding Member of the Academy, and Academicians A.Ye. Fersman and F.Yu. Levinson-Lessing, on the exchange of scientific material.[5]

Soviet geologists, highly appreciative of the effort of their Chinese colleagues,[6] attached great significance to scientific cooperation in the study of the geology of eastern Siberia and China. A meeting of the Section of Physical and Mathematical Sciences of the Academy of Sciences voted approval to Li Sy-Huan's proposition, as presented by A.Ye. Fersman, of a joint expedition to the Baikal region and Manchuria.

At the same time, Li Sy-Huan became well acquainted with the Mineralogical Museum of the Academy. In his interview with the Izvestiya correspondent, he said, "I am tremendously impressed with what I have seen in your museums. As a geologist, I am very much interested in your rich collection of gems, pearls, platinum, gold, and other specimens. This is the first time I have seen such scientific wealth."[7]

Li Sy-Huan took advantage of his visit to the capital, by getting acquainted with the geology of the Moscow trough. Accompanied by M.A. Bolkhovitina, he made a trip to Podol'sk where he collected the material he needed. Comparing his own work with that of M.A. Bolkhovitina, Li Sy-Huan was able to note the great similarity between Middle Carboniferous marine deposits of China and the U.S.S.R.

In his letter of May 28, 1958, Li Sy-Huan reminisces as follows, "Upon my return from the Soviet Union, in 1958, I reported to the Chinese Geological Society. As far as I can remember, the gist of it was 1) the warm sympathy of the Soviet people for the people of China, and the reception given to me — its representative — at the meeting in a Moscow theater; 2) that the toilers indeed have been liberated under Soviet power, and their children have priority in entering the university; also, that science has ceased to be the monopoly of a handful of the bourgeois class but has become the property of the working masses; 3) in the field of geology, innovations have been introduced into the old system and a reorganization has been carried out; it became clear that under the leadership of the Bolshevik Party and the Soviet Government, geologic science will achieve a tremendous development in the near future. Evidence of that is already here."

Personal contacts, established by Li Sy-Huan with Soviet Scientists in Moscow and Leningrad have been maintained. A.N. Krishtofovich and P.I. Polevoy visited China. In 1926, Academician V.L. Komarov, Corresponding Member B.Ya. Vladimirtsev, and L.S. Berg, a Member of the Commission for the Study of Natural Productive Potentiality of the U.S.S.R. — subsequently Academician — also went to China.

The scientists of both countries exchanged scientific literature. The Soviet students of Far Eastern geology kept in touch with the Chinese Geological Society.

With the establishment, in 1927, of the reactionary dictatorship of the Kuomintang, the conditions for Sino-Soviet scientific collaboration deteriorated markedly. However, the scientists of both countries did all they could to maintain their relations. For instance, Li Sy-Huan was instrumental in having the A. Semikhatov work on Carboniferous stratigraphy published in the bulletin of the Geological Society of China. Outstanding Soviet geologists were elected to that Society.

The cooperation between Chinese and Soviet geologists blossomed forth after the proclamation of the Chinese People's Republic. This cooperation has been promoted to a considerable extent by Li Sy-Huan, now Minister of Geology, C.P.R., Vice-President of the Academy of Sciences, C.P.R., and Chairman of the All-China Society of Natural Sciences.

Speaking before the first session of the Second People's Consultative Council of China (April, 1959), Li Sy-Huan drew a bright picture of the "Big Leap" in the field of geology. The facts which he has cited are convincing proof of the success of Chinese geology.

[4] Letter of Li Sy-Huan to the author, April 5, 1958 (Original in Russian).

[5] Letter of Li Sy-Huan to the author, May 28, 1958.

[6] See A.Ye. Fersman, Izbrannyye Trudy, t. 1, Izd. Akademiya Nauk SSSR, 1958, p. 368.

[7] Izvestiya (newspaper), September 11, 1925.

THE 1959 AWARD OF THE A.P. KARPINSKY GOLD MEDAL[8]

By a resolution of the Presidium of the Academy of Sciences, U.S.S.R., May 29, 1959, the 1958 A.P. Karpinskiy gold medal was awarded to an outstanding scientist, Vice-President of the Academy of Sciences of the Chinese People's Republic, Vice-President of the World's Federation of Scientists, Foreign Member of the Academy of Sciences, U.S.S.R., Li Sy-Huan, for his scientific work in the field of Geology, Paleontology, Petrography and Minerals.

[8]O prisuzhdenii zolotoy medali imeni A.P. Karpinskogo 1958 g.

REVIEWS AND DISCUSSIONS

A FEW OBSERVATIONS ON THE BOOK OF S.G. SARKISYAN, "MESOZOIC AND TERTIARY DEPOSITS OF THE BAIKAL AND TRANS-BAIKAL REGIONS AND THE FAR EAST"[1,2]

Recent years have witnessed fairly extensive geologic surveying and reconnaissance work done in the Trans-Baikal region and the Far East, by a large number of geologists from both local and central geologic organizations of the Ministry of Geology and Conservation of Mineral Resources, and of the Academy of Sciences, U.S.S.R. and its affiliates. Unfortunately, there still is no general work on the geology of that vast region. The book by S.G. Sarkisyan has not filled the gap.

The main shortcomings of this book are the lack of completeness in its material, the large number of errors in the presentation of data, the lack of correspondence between the stratigraphic descriptions and the present status of knowledge, and the inadmissibly large number of errata, some of which are undoubtedly due to the author's negligence.

The incompleteness of material used. As stated in the introduction, this book is a compilation of voluminous published material. It appears from the bibliography (pp. 330-337) that the author has used some of the recent 1957 and 1958 publications, such as works of M.V. Korzh and N.N. Sokolova. Therefore, the reader is fully justified in expecting this work to reflect at least the publications prior to the second half of 1957. However, the author has generally not made use of extensive material on that region, published between 1953 and 1956, not even that of the Conference on the Unified Stratigraphic Classification of the Far East, the abstracts of whose reports and papers were published in 1956, with the originals kept in the libraries of geologic organizations of Moscow, Leningrad, Vladivostok, and Khabarovsk. It is true that the author refers to the resolutions of that Conference in his description of Triassic sections (p. 55) but they are neither considered nor mentioned anywhere else.

No credit at all is given to the geologic study in Sikhote-Alin by the members of the Fourth and the Far Eastern Geological Administrations and of the All-Union Geological Institute (VSEGEI): Ye.B. Bel'tenev, A.I. Savchenko, and others (Doklady Akademiya Nauk SSSR, 1956, 110, no. 5), L.I. Krasnyy (Inform. Sborn. VSEGEI, 1956, no. 3), and many others. Without these works, one cannot gain a proper understanding of the Mesozoic stratigraphy of northern Sikhote-Alin and the coast of the Okhotsk Sea.

For reasons difficult to understand, the Triassic is described only for the southern Maritime Province, while the faunally-described Triassic deposits of middle and northern Sikhote-Alin, Dusse-Alin, the Okhotsk Sea coast, and the adjacent regions of Tyrma River, Upper Amur, and the Trans-Baikal, are not even mentioned. Thus, the S.G. Sarkisyan book gives the reader a very distorted impression of the distribution of Triassic deposits and of their significance in the geologic history of that region.

The chapter, "Khabarovsk Trough," is written from the obsolescent pre-war data of I.G. Kozlov; a stratigraphic section shows as Jurassic, deposits whose age has been faunally determined as Cretaceous. This is because the author has overlooked the new work of the All-Union Aero-Geological Trust (VAGT), VSEGEI, and the Fourth and the Far Eastern Geological Administrations, in the period of 1953 to 1957.

Errors in the presentation of material. We confine ourselves mostly to the errors in the presentation of material on Jurassic deposits which we have studied in the last few years.

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[1]Nekotoryye zamechaniya k knige S.G. Sarkisyana "Mezozoyskiye i tretichnyye otlozheniya Pribaykal'ya, zabaykal'ya i Dal'nego Vostoka."

[2]Academy of Sciences, U.S.S.R., publication, 1958.

The Jurassic section in Sikhote Alin, including Southern Maritime province

	Ammonites	Lamellibranchia and other	Composition of deposits and their distribution
Upper and middle Tithonian	Berriasella sp.		Sandstone of Putyatin Island and Abrek Peninsula
Lower Tithonian	Primoryites primoryensis, Virgatosphinctes contiguus, Aulacosphunctes (Torquatisphinctes?) sp., Subplanites sp. and other	Trigonia e.g. formosa, Tr. ivantischini, Tr. e.g. ivantischini, Tr. aff. doroscheni, Pinna subradiata, Camptonectes e.g. cinctus, Variamussium nonarium Astarte and other	Mostly sandstone, with some conglomerate, siltstone, and limestone. The deposits occur on Askol'd and Putyatin islands; along the Ussuri Bay coast; in basins of Shitukha (Galanta River), Taudema, Rakovka, Suputnik, Sydagou Rivers, and in the upper course of the Bikin (Tavasichka, Chinga Rivers), and Valinok River (a tributary of Iman)
Kimmeridgian	Rasenia sp.	Aucella e.g. mosquensis Aucella e.g. bronni, Inoceramus sp.	Mostly sandstone with thin intercalations of siltstone, developed along the Ussuri Bay shore (Chagan Point); in the area of Linda railroad station; in quarries along the highway from Linda to Promyslovka; in the valleys of Taudema, Suchan, upper Bikin (Chinga R.) Rivers; in the area of Nakhodka Bay and Abrek Peninsula. Conglomerate, sandstone, and siltstone along the Ussuri Bay coast and in the lower Bikin area
Oxfordian			
Callovian	Fauna not discovered		
Bathonian			
Bajocian			
Aalenian	Holcophylloceras ussuriensis	Inoceramus formosulus, In. ussuriensis, In. lucifer, In. subambiguus and other	Mostly sandstone in the area of Suyfun, Nants, and lower Bikin Rivers
Toarcian	Grammoceras sp., Lillia sp.		Sandstone and siltstone in the basin of Sydagou and Khungari Rivers
Domerian Pliensbachian	Uptonia sp. Acanthopleuroceras (?) sp.	Oxytoma cygnipes, Pecten cf. riniki, P. textorias, Harpax laevigata, H. senecseks, Cardinia sp. and other	Gravel and sandstone, locally tuffaceous, in the Shitukha basin; siltstone and sandstone in the upper Bikin basin
Lotharingian Sinemurian	Arnioceras sp.		Conglomerate, gravel, and sandstone of the Sydagou basin
Hettangian	Fauna not discovered		

Pages 77 and 78 carry a description of the "Upper Mongigay formation" and the "Noric stage" throughout the basin of Iman River and its tributaries. As a matter of fact, S.G. Sarkisyan is describing here the Aptian-Albian sections characterized by a fauna described by I.V. Buriy, in 1956 (Aucellina aff. aptiensis Pomp., etc., as identified by V.N. Vereshchagin and Yu.G. Mirolyubov).

S.G. Sarkisyan's diagrammatic section of Jurassic deposits from the southern part of the southern Maritime Province (Table 5, p. 94) does not correspond to nature. The work of geologists of the Fourth Geological Administration (B.I. Vasil'yev and others), in 1955 and 1956, has established that Trigonia, assigned by S.G. Sarkisyan to the Bajocian and Callovian, occurs in reality above the Aucella sandstones — Lower Volgian, according to S.G. Sarkisyan — while his "Oxfordian" fauna occurs above the "Bajocian-Callovian" and "Lower Volgian."

Deposits described in pp. 96-106 (down to paragraph seven of Chapter "The Basin of Kolumbe and Nants Rivers") can be assigned to the Jurassic only very tentatively, because of the Lower and Upper Cretaceous deposits widely developed in that area.

Also at variance with the true situation is the author's statement (p. 107) on the distribution of Middle Jurassic deposits along the Ussuri Bay coast, and farther on, in an unbroken belt from the Pashkeyev Falls, southwest as far as Galanta River, and farther on southeast; also on the presence of isolated Middle Jurassic outcrops at Chagan Point, in the Pod"yampol'skiy Bay, and the island of Putyatin. At all those places, with the exception of the Pod"yampol'skiy Bay, a fauna was collected of Upper Jurassic age of (ammonites Virgatosphinctes sp., Aulacosphintes sp., etc.).

The book contains a "Generalized Paleogeographic Map for Lower and Middle Jurassic Time, in the Trans-Baikal Region and the Far East." (Fig. 16). This, however, is not a paleogeographic map, inasmuch as it does not show either land or sea.

There are direct contradictions in the book. Thus on page 94, the Chagan Point sandstone is assigned to the Kimeridgian; it is Middle Jurassic on page 107, and Volgian on page 113. The Putyatin Island sandstone is Middle Jurassic on page 107, and Upper Jurassic on page 113.

The stratigraphic table on page 129 does not fit the text. In the table, the Umal'tin formation is strictly Lower Jurassic, while page 130 ascribes to it a fauna (Lioceras cf. brasile) typical of the Middle Jurassic. The Elgin formation is assigned to the Upper Jurassic, in both the table and the heading on page 130; page 131 refers to it as Bathonian-Callovian, i.e., Middle to Upper Jurassic.

We could go on enumerating more contradictions and inconsistencies with the present status of knowledge. We will not, however, take any more of the reader's time, except for presenting the latest classification of the Jurassic section in Sikhote-Alin, including the southern Maritime Province section (see Table on page 109, this magazine).

A comparison of this classification with Table 5, page 94 of the book under review leads to the following conclusions:

1. Jurassic stratigraphy of Sikhote-Alin can be constructed from the ammonite fauna, and in much greater detail than has been done by S.G. Sarkisyan.

2. The presence of the Hettangian stage (Lower Jurassic) has not been faunally proved for Sikhote-Alin.

3. There also is no faunal proof of the presence of the Bajocian, Bathonian, and Callovian, which are cited by S.G. Sarkisyan. This is especially important, inasmuch as he ascribes a great importance to the Middle Jurassic, in estimating its thickness at 2,200 meters (page 162).

Errata. They are numerous in the book. Thus in the description of the Jurassic, page 94, Table 5, we read Idiocyloceras sp. where it should be Idiocycloceras sp.; on p. 96, Armioceras sp. instead of Arnioceras sp.; on p. 107, Acantoplearoceras sp. indet. instead of Acantopleuroceras sp. indet.; on p. 113, Vdiocyloceras Spath instead of Idiocycloceras Spath, and Pelaceroides instead of Peltoceradoides. On pp. 113-114, Aucella broni is given four times instead of Aucella bronni; on p. 115, Nucula oxfordina karenia sp., Lyticeras sp., instead of Nuculus oxfordiana, Rasenia sp., Lytoceras sp.; on p. 116, Rhylloceras sp. indet. instead of Phylloceras sp. indet.; on p. 119, there is a mysterious inscription, Astrate-Striata-Costa. There are similar mistakes nearly everywhere throughout the book wherever there are lists of flora and fauna. There are also other "errata;" for instance, a fictitious "Katen-Iman watershed" is described on page 89 (see geographic map of Sikhote-Alin). The last five paragraphs on page 106 are erroneously included in the chapter on "Basins of Kolumbe and Nants Rivers," although they describe sections along the lower course of Bikin River and should therefore be placed at the end of the next chapter.

Described on page 115, in the chapter on "The Iman River Basin" are Jurassic sections developed along the lower Bikin course, and in northern areas, which have nothing to do with the Iman basin. Table 8, pp. 138-140, cites a "Mesozoic system" twice; and the "Middle Jurassic" and "Upper Jurassic" are designated as two formations of the Lower Cretaceous (see p. 140).

The author probably is well aware that there is no "Mesozoic system," and that the Middle and Upper Jurassic are not Lower Cretaceous formations. We cannot gloss over such carelessness. We feel justified in demanding from an author and an editor a serious view of their work and due regard for the reader — especially a young reader who, getting hold of a book published on the authority of the Academy of Sciences, sincerely believes in its scientific reliability. There should be but one opinion of the "usefulness" of such a book.

ON THE SO-CALLED PRE-TEL'POS ORDOVICIAN IN THE POLAR URALS[3]

The Bulletin of Scientific-Technical Information of the Ministry of Geology and Conservation of Mineral Resources of the U.S.S.R., no. 4(16), 1958, carries a communication of A.P. Belousov of finding "pre-Tel'pes" Ordovician in the Polar Urals. This communication cannot but interest the geologists who work in that region and all those who have anything to do with Ordovician stratigraphy. The Tel'pes formation, described from the Near-Polar Urals by K.A. L'vov [3, 4], is a synonym for the Miniseysk formation of the Polar Urals [5]. This latter formation not only carries a number of Tremadoc and Arenigian forms, such as Angarella cf. obritschewi Assatk., A. ex. gr. jawoskii, etc., but also representatives of genus Billingsella, typical throughout the Upper Cambrian.

For that reason, the presence of "pre-Tel'pos," or more precisely pre-Miniseysk Ordovician would be of great academic interest. Its existence would be regarded as certain if the Khuutin rocks with an Ordovician fauna had been observed below the basal Miniseysk (Tel'pos) horizon. This would make it necessary to reconsider the upper limit of representatives of genus Billingsella.

Unfortunately, the A.P. Belousov communication does not contain any indication of such relationship in the Polar Urals. The data cited only establish the fact that in a number of localities along the upper courses of Bol'shaya Usa and Bol'shaya Paypudyna Rivers and northeast of lakes Bol'shoye Khadata — Yugan-Lor, the "Khuutin banded schist" lie directly under carbonate rocks with a Silurian fauna or else alternate with "Upper Ordovician Shchugor limestone." Such relationship means only that the "banded schist" does not belong to the Khuutin formation alone and cannot be a positive criterion of the latter's presence in an area. Accordingly, a more correct title for A.P. Belousov's communication should have been, "On the Presence of Banded Schist in the Ordovician of the Polar Urals." This would have been in better accord with the data cited.

Therefore, the oldest known Ordovician deposits in the Polar Urals still are the Tel'pes formation and its equivalents. The subject of the stratigraphic distribution of Billingsella, too, does not have to be reopened, for the time being.

[3]O tak nazyvayemom "Dotel'posskom ordovike" na polyarom Urale.

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